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(Continued on inside back cover)

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INSOLATION AS AN EMPIRICAL FUNCTION OF DAILY SUNSHINE DURATION

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ABSTRACT

Empirical relations between incident solar radiation received at the earth's surface and (1) percent of possible sunshine, (2) latitude, and (3) time of year are developed. These relations are combined into a graphical method for converting percent of possible sunshine into daily values of incident solar radiation for stations between latitudes 25° N. and 50° N. The method is tested on independent data from widely separated locations and a correlation coefficient of 0.97 between estimated and observed values is obtained.

INTRODUCTION

Easily obtained estimates of the daily amount of incident solar radiation (insolation) received at the earth's surface, where observed values are unobtainable, have practicable application in the field of applied meteorology and hydrology. Such values have an immediate utility in the energy balance method of estimating heat and vapor transfer at snow, water, soil, or plant surfaces. This paper presents an empirical-graphical method of converting observed values of the percentage possible hours of sunshine into estimates of insolation.

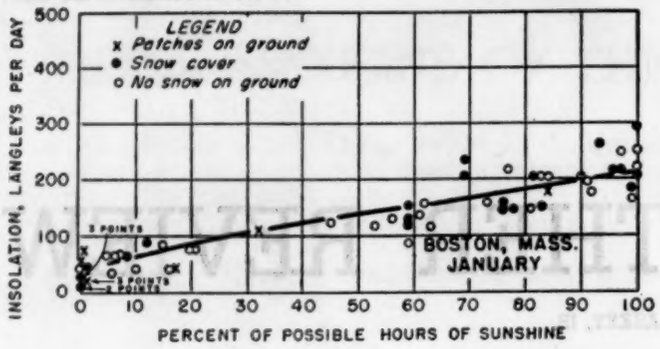
The network of pyrheliometer stations in the United States has expanded during recent years but the number of observations is not sufficient to define the areal distribution of insolation for short periods. There were 49 such stations in 1953 in the continental United States operated by the Weather Bureau, and 21 cooperative stations. A conversion of the percent possible sunshine obtainable from approximately 170 Weather Bureau stations into estimates of insolation would greatly improve the accuracy of the interpolated areal distribution.

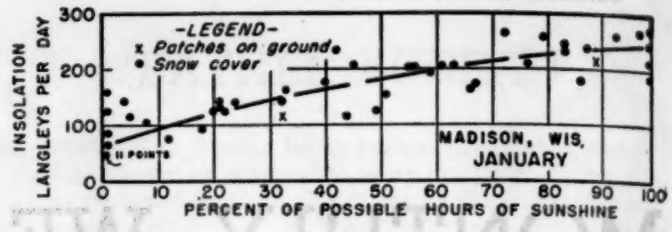
Values of average theoretical direct solar radiation reaching the ground under cloudless conditions are available in the Smithsonian Meteorological Tables [1]. These computed values consider the following parameters: (a) solar constant, (b) radius vector of the earth, (c) zenith distance of the sun, (d) a transmission coefficient for the atmosphere. The solar constant is considered to have a value of 1.94 ly/min and the zenith distance is a function of the latitude of the station, declination of the sun, and hour angle. For a particular station the quantity of solar radiation that is transmitted to the earth's surface, direct and diffuse, is a complicated function involving station elevation, character and amount of cloudiness, water vapor, kind and amount of pollutants—dust, smoke, etc. The percent of possible hours of sunshine, as determined by the sunshine recorder at the station, has been used as an indicator of the combined effect of these variables.

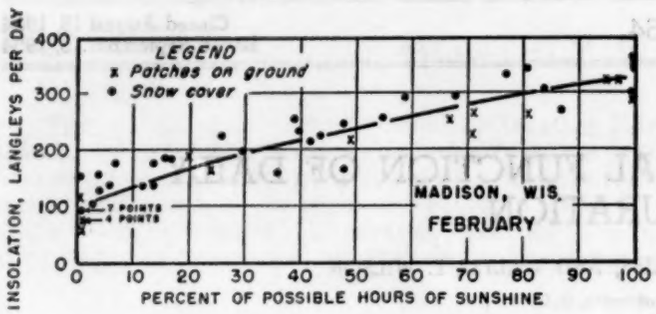
TABLE 1.—*List of stations from which observations were obtained*

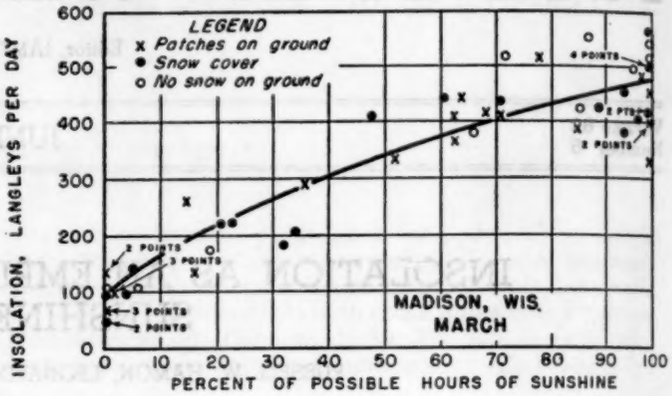
Station	Lat. ° N	Station	Lat. ° N
Miami, Fla.	25.8	Washington, D. C.	38.8
Brownsville, Tex.	25.9	Columbia, Mo.	38.9
San Antonio, Tex.	29.5	Indianapolis, Ind.	39.7
Appalachicola, Fla.	29.7	Lincoln, Nebr.	40.9
Charleston, S. C.	32.8	Cleveland, Ohio	41.3
Atlanta, Ga.	33.5	Boston, Mass.	42.4
Little Rock, Ark.	34.7	East Lansing, Mich.	42.8
Oklahoma City, Okla.	35.4	Madison, Wis.	43.1
Greensboro, N. C.	36.0	Sault Ste. Marie, Mich.	46.4
Nashville, Tenn.	36.0	Bismarck, N. Dak.	46.8

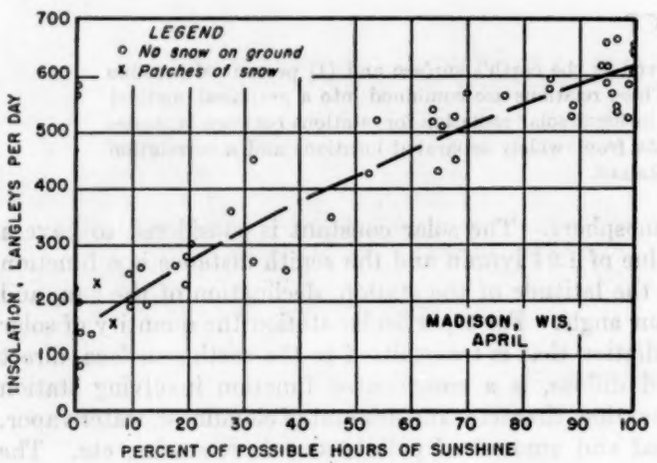
For use in developing the empirical method observations of percent possible sunshine S, insolation Q, and snow on the ground, for the years 1951, 1952, and 1953 were obtained from Weather Bureau records [2] for 20 stations, listed in table 1, in the continental United States.

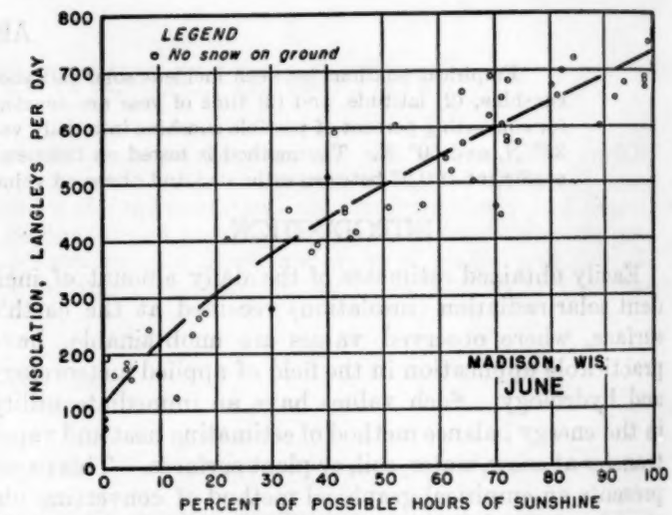


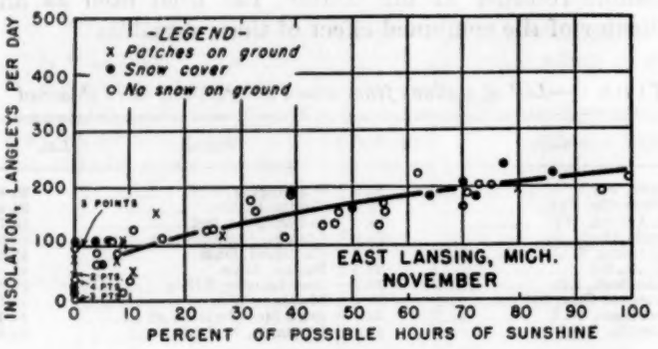












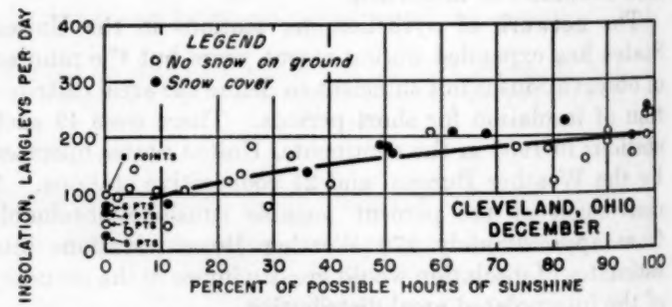


FIGURE 1.—Representative scatter diagrams showing observed insolation as a function of observed percent of possible duration of sunshine, daily values.

DEVELOPMENT OF METHOD

The relation between insolation and the percent of possible sunshine was first investigated by plotting scatter diagrams for each station and month and fitting curves to the points by inspection. Figure 1 displays representative curves that resulted; the curvilinear relation corresponds to that obtained by Kimball [3]. The existence of snow cover has been reported [4, 5] to produce an increase in the diffuse radiation for overcast conditions. To determine the magnitude of this effect the points on the scatter diagrams were identified for those days having snow on the ground, patches of snow, or no snow. The scatter diagrams failed to show sufficient separation of the points to enable separate curves to be drawn to snow and no-snow cases; therefore they were considered as the same population in defining the curves. The values of Q for each station at 100, 80, 60, 40, 20, and 0 percent possible hours of sunshine were read from the curves of the scatter diagrams and Q/Q_0 determined for each month for each station, where Q_0 is the value read from the curve at 100 percent possible hours of sunshine.

In a study of the relation between average monthly values of Q and S, Fritz and MacDonald [6] found that for eleven stations between latitudes 25° and 44° N, a plot of Q/Q_0 versus S could be fitted by a straight line $Q/Q_0=0.35+0.61S$. Their data contained S values ranging from 0.35 to 0.97. The correlation coefficient obtained was 0.88. Their work referred only to monthly average values of Q and S and is, therefore, essentially different from the work described in this paper. Similarly, Black, Bonython, and Prescott [7], using monthly mean data for 32 stations give the relation as $Q/Q_0=0.23+0.48\,S$.

It has been suggested [8, 9, 10] that the relation between Q/Q_0 and S, for daily values is

$$Q/Q_0=k+S(1-k) \qquad (1)$$

where the reduction factor k is the ratio of total radiation with zero percent sunshine to total radiation with 100 percent sunshine. A refinement of the above relation may be represented by

$$Q/Q_0=k+C_s(1-k) \qquad (2)$$

where C_s, the variable sunshine factor, is a function of S. The subscript s refers to the value of the factor at S percent sunshine. From the observed values of S and the values of Q/Q_0 determined above, C_s was found empirically to be the same for all latitudes for a particular value of percent sunshine. The empirical relation of C_s to S is shown in figure 2. The circles are the mean values used to define the curve. On the other hand, k was found to vary slightly with season and considerably with latitude. The latitude variation is illustrated in figure 3 for mean annual values for the stations in table 1. The curve of figure 3 was fitted to the points by inspection. A k value of .40 for Fairbanks, Alaska, latitude 64.8° N. is not shown in the figure, but it was considered, as were some higher latitude values given by Kalitin [5], in determining the shape of the curve. The seasonal variation of k is shown in the table insert of figure 5 (zero correction with 80 or more percent sunshine).

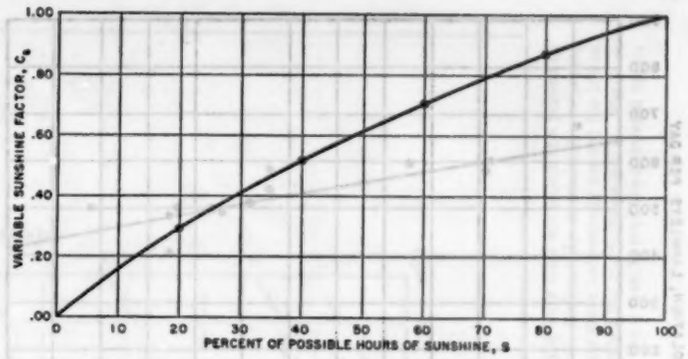


FIGURE 2.—Relation of the variable sunshine factor C_s to percent sunshine S.

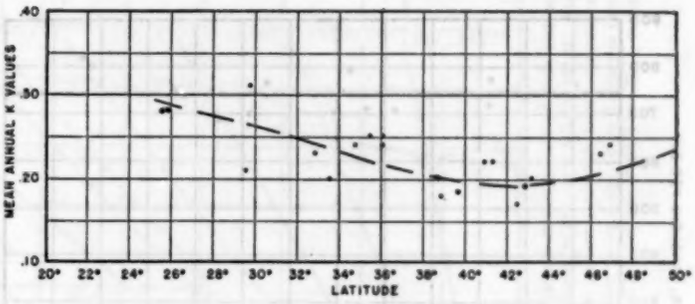


FIGURE 3.—Relation of mean annual values of Q/Q_0 for zero percent sunshine (i. e., k) as a function of latitude for stations listed in table 1.

The 100 percent sunshine data for March, June, September, and December for each station were plotted separately versus the station latitude in figure 4, and compared with calculated values of Q_0 as obtained from the Smithsonian Tables [1]. The curves in figure 4 show the calculated values, made to fit the empirical data by appropriate choice of atmospheric transmission coefficients. The following coefficients were derived by trial and error: March 0.80, June 0.70, September 0.75, and December 0.85. The coefficients for the intermediate months can be interpolated. From these curves the curves labeled "latitude" of figure 5 were constructed. Kennedy [11], uses the formula $I=I_0a^m$ relating the insolation at the ground, I, to that exterior to the atmosphere I_0, the atmospheric transmission coefficient, a, and the solar air mass, m. From 2 years record at Fresno, Calif., and Lincoln, Nebr., he finds the transmission coefficient for clear days to be 0.91. Gerdel, Diamond, and Walsh [12], use this formula and assume an atmospheric transmission coefficient of 0.90 for all seasons to compute a set of "latitude" curves. These curves give around 50 ly/day less insolation than does figure 5.

As shown in figure 4, the Q values with 100 percent of possible sunshine vary only slightly with latitude at the summer solstice, approximating 740 langleys per day.

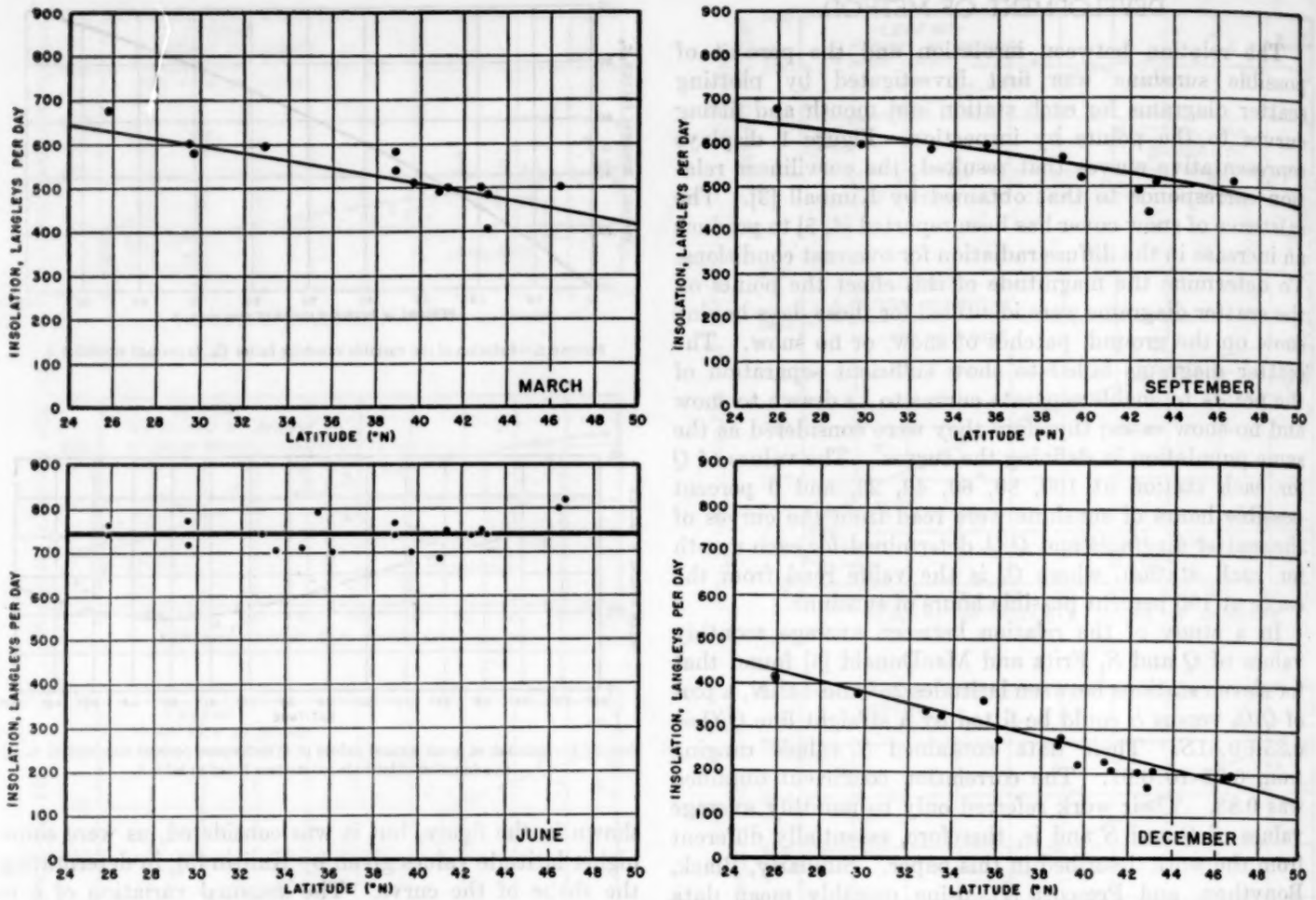


FIGURE 4.—Observed insolation for days having 100 percent sunshine, as a function of latitude and mean monthly values, during March, June, September, and December, for selected stations from list of table 1.

This is mostly because the increase in daylight hours with latitude counterbalances the decrease in solar altitude. The curve of k as a function of latitude, figure 3, is the curve for Q/Q_0 for zero percent sunshine in figure 5. These values of k together with those of C_s from figure 2 were used in equation (2) to determine the reduction factor variation with latitude for the remaining curves of the other percent sunshine values shown in the right side of figure 5.

The empirical relations between insolation Q received at the earth's surface and (1) percent of possible sunshine S, (2) latitude, and (3) time of year thus have been combined graphically in figure 5 to provide a working chart for estimating daily values of insolation.

TEST OF METHOD

The empirical relationship of S and Q as developed has been tested on independent data from widely separated locations: Salt Lake City, Utah, 1952; Seattle, Wash., 1952; Madison, Wis., 1950; Atlanta, Ga., 1953; Appalachicola, Fla., 1953; and Portland, Maine, 1950. For this test, data used were for the 1st, 10th, and 20th day of each month of the year. For a few of the days these data were missing. A total of 207 cases were estimated from the graph of figure 5. As an evaluation of the estimates obtained from the relation, the usual correlation test was made of values estimated from the graph, versus observed values. These are plotted in figure 6. The correlation coefficient obtained was 0.97, with a standard error of estimate of 36 langleys on a daily basis. This corresponds to about 170 langleys on a monthly basis. (If seasonal trend is removed the correlation coefficient drops to 0.84.)

Much of the residual scatter of the estimated insolation values results from the variability in the character of clouds and other restricting phenomena, and their time of occurrence. Also, for a particular station the Q estimated from the S data may show a systematic difference from the observed Q values due to local or regional diminution of radiation attributable to factors such as haze and industrial pollutants in the atmosphere. Station correction factors would produce some improvement and could be obtained for all stations with Q observations, and isolines constructed for adjustment in the values of Q obtained from the empirical method [10]. Difficulty would arise, however, in interpolating S values to dis-

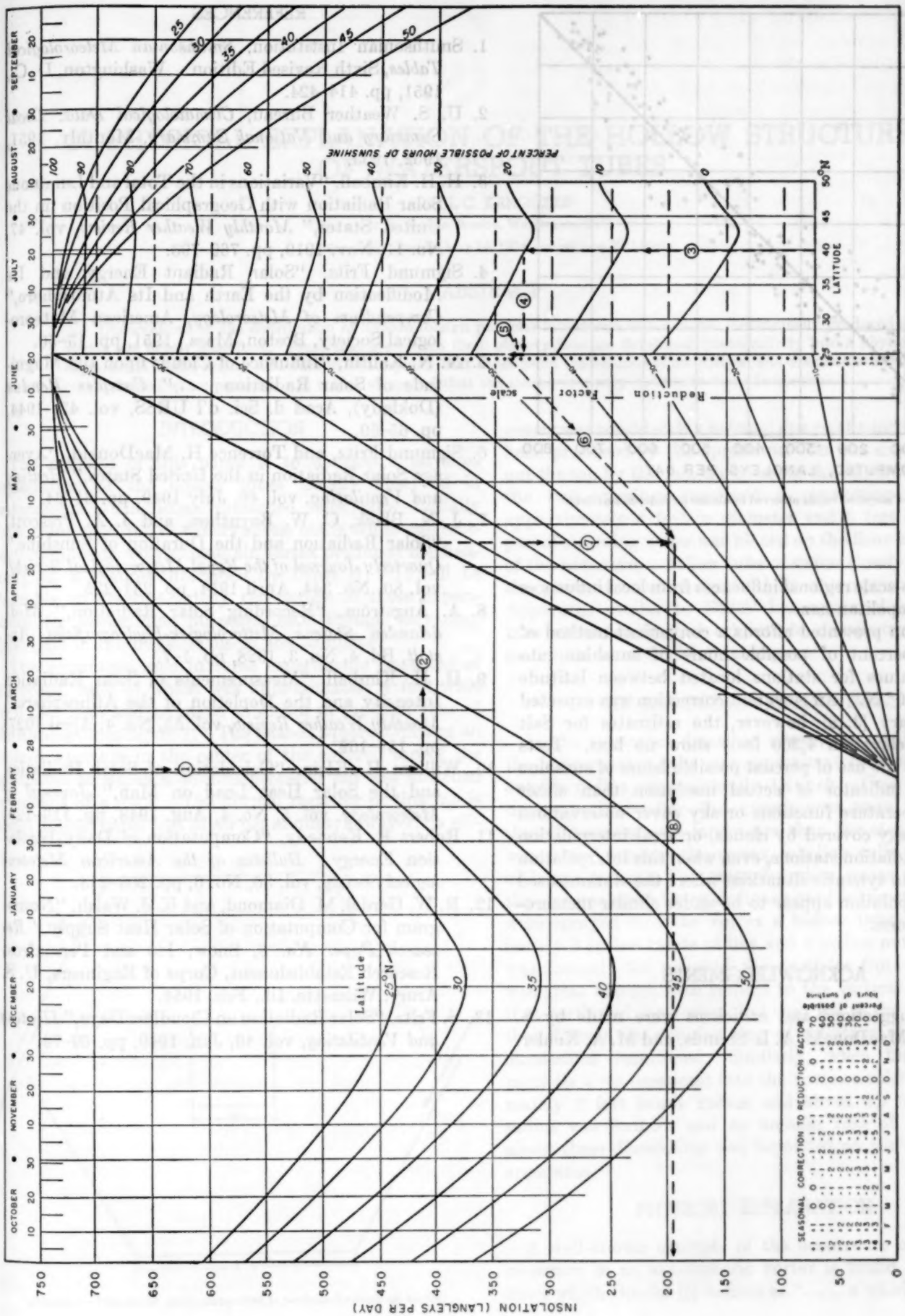


FIGURE 5.—Working diagram for computing insolation as a function of percent possible sunshine, latitude, and time of year. The example shown by dashed lines with arrows estimates the daily total insolation at a station at latitude 40° N., on February 21 and with 20 percent possible sunshine: Step 1.—enter the graph at February 21 and move downward, to the curve labeled 40° N. Step 2.—thence move horizontally to establish a reference point at the intersection with the heavy line. Step 3.—enter lower right of the graph at latitude 40° N., proceeding upward to 20 percent of possible hours of sunshine. Step 4.—thence move horizontally toward the left to the vertical line through latitude 25° N. Step 5.—here the seasonal correction (+2) from the insert table is added on the reduction factor scale. Step 6.—thence move downward toward the convergence point. Step 7.—from the reference point of step 2, move downward to an intersection with the line extended in step 6. Step 8.—from the point of intersection of these two lines, move horizontally to the Q scale to read the estimated value, 190 langleys per day.

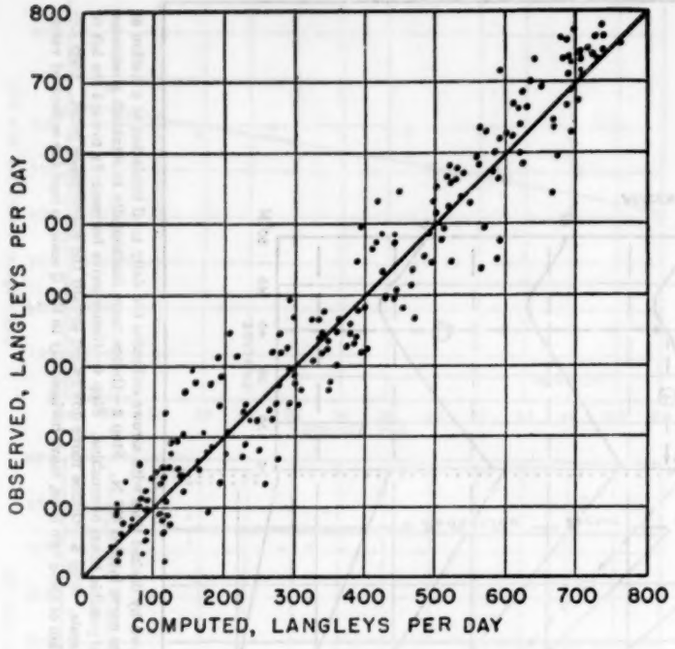


FIGURE 6.—Computed versus observed insolation for independent data.

tinguish large-scale regional influences from local influences such as metropolitan haze.

The relation presented affords a convenient method of converting percent of possible hours of sunshine into insolation values for stations located between latitude 25° N. and 50° N. An elevation correction was expected to be necessary [3, 8]; however, the estimates for Salt Lake City (elevation 4,260 feet) show no bias. Tests indicate that the use of percent possible hours of sunshine is a better indicator of actual insolation than either diurnal temperature functions or sky cover observations in tenths of sky covered by clouds, or areal interpolation among the radiation stations, even when this interpolation is restricted to synoptic situations where the stations used for the interpolation appear to be under similar meteorological influences.

ACKNOWLEDGMENTS

Helpful suggestions and criticisms were made by S. Fritz, T. H. MacDonald, A. L. Shands, and M. A. Kohler.

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A PHYSICAL EXPLANATION OF THE HOLLOW STRUCTURE OF WATERSPOUT TUBES

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[Manuscript received May 14, 1954; revised June 29, 1954]

ABSTRACT

Observations of the distribution of minute foreign particles introduced into a small, mechanically produced air vortex are briefly described. Ideas that grew from these observations are developed quantitatively into a physical explanation of the hollow structure of waterspouts that depends on the balance of centrifugal and drag forces acting on condensed water particles. It is also suggested that the mechanism may operate in tornado vortices.

INTRODUCTION

The purpose of this paper is to develop a physical explanation of the hollow tubes usually observed in waterspouts. The ideas presented here grew initially from some qualitative observations made while conducting experiments with a small air vortex produced mechanically in a manner described by Dines [1]. It will be desirable to describe briefly the nature of these experiments for background purposes, even though they play little part in the physical discussion to follow.

The apparatus used in these experiments consisted of 6 screens, each of which was approximately 4 feet wide by 6 feet high and arranged as shown in the plan view of the apparatus in figure 1. The dashed square indicates an 18-inch aperture in the center of the covered top of the apparatus to which a 36-inch propeller fan was attached by means of a duct. The normal capacity of this fan is about 10,000 cubic feet of air per minute; however, its efficiency was considerably reduced due to the increased static head produced by the apparatus. As the air was sucked out the top by the fan, air entering the apparatus through the two openings produced a converging air vortex approximately 8 feet in diameter and 6 feet high. If a pan of steaming water was placed on the floor in the center of the enclosure, a hollow tube of vapor closely resembling the pendant column observed in waterspouts appeared. Apparatus similar to Dines' has appeared from time to time at meteorological exhibits to demonstrate the formation of an artificial "waterspout."

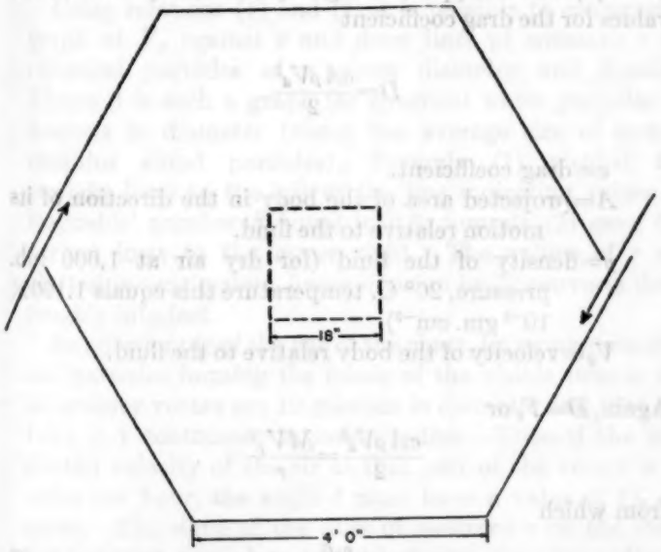

FIGURE 1.—Plan sketch of apparatus used to produce the small air vortex.

EXPERIMENTAL OBSERVATIONS

In the apparatus described above, the inside radius of the hollow steam tube was approximately ½ inch and the outside radius approximately 1½ inch. One of the first things noticed after the experiments were begun was that if particles of a larger size than the steam particles were broadcast into the vortex, a hollow tube was formed of larger radius. For example, when water particles of the size produced by an ordinary garden insecticide spray were sprayed into the vortex a hollow tube of approximately 3 inches inside radius and 5 inches outside radius was formed; the mass of the particles from this spray was great enough with respect to the vertical velocity in that portion of the vortex that they fell to the floor where they formed an annular ring of moisture of about the dimensions mentioned. Similarly, when fine sawdust particles were broadcast into the vortex a tube of approximately 2 feet inside radius and about 2½ feet outside radius was formed, and an annular ring of sawdust of about these dimensions was deposited on the floor of the apparatus.

PHYSICAL EXPLANATION

A well-known example of the occurrence of this phenomenon in an atmospheric vortex is found in the dust devil which Brooks [2] defines as ". . . a whirling column

of dust or sand, in which the size of the particles increases with the distance from the center." It is proposed here that this phenomenon occurs because each foreign particle [1] introduced into the vortex seeks an equilibrium distance from the vortex axis where the inward force due to the frictional drag on the particle caused by a converging flow of air is balanced by the outward force on the particle due to centrifugal force. It should be evident that when reaching this equilibrium position, a spherical particle in a stable, non-turbulent vortex will travel in a circular path with constant velocity.

Let us begin by considering such a particle moving at constant speed in a circular path in such a vortex. The forces acting on the particle are illustrated in figure 2, where

V = velocity of the medium (air)
V_p = velocity of the particle in its path
V_d = vector difference between V and V_p
F_c = centrifugal force on the particle due to its motion
D = frictional force or "drag" produced on the particle by the motion of the medium relative to the particle
θ = angle between V and V_p.

It will be noted that the aerodynamic force on the particle caused by the wind shear in the vortex has been omitted. The order of magnitude of this force is such that it may be neglected for our purposes.

The drag force D acts in the direction of the vector expressing the velocity of the medium with respect to the particle, or along the vector V_d which expresses the vector difference between the velocity of the air and the velocity of the particle in its motion around the circular path; both V_d and D are directed toward the center of the circle of motion, and $D=F_c$.

Experiments made to determine the frictional drag of spheres in air [3] indicate that for Reynolds' numbers up to about 0.5, Stokes' formula for the frictional drag D of a sphere of diameter a, moving through an incompressible fluid of viscosity μ, with a velocity V_d with respect to the fluid, gives good agreement with the experimental values. This formula

$$D=3\pi\mu a V_d,$$

shall be used for Reynolds' numbers up to 0.5 and experimental values of the drag coefficient [3] shall be used for Reynolds' numbers greater than 0.5.

Now for a particle of mass M

$$F_c=\frac{MV_p^2}{r},$$

where r=radius of the path. When the particle is travelling at constant speed in a circular path, $D=F_c$, or

$$3\pi\mu a V_d=\frac{MV_p^2}{r}.$$

[1] The word "particle" throughout this discussion will always signify a foreign particle introduced into the vortex as distinct from a parcel of air.

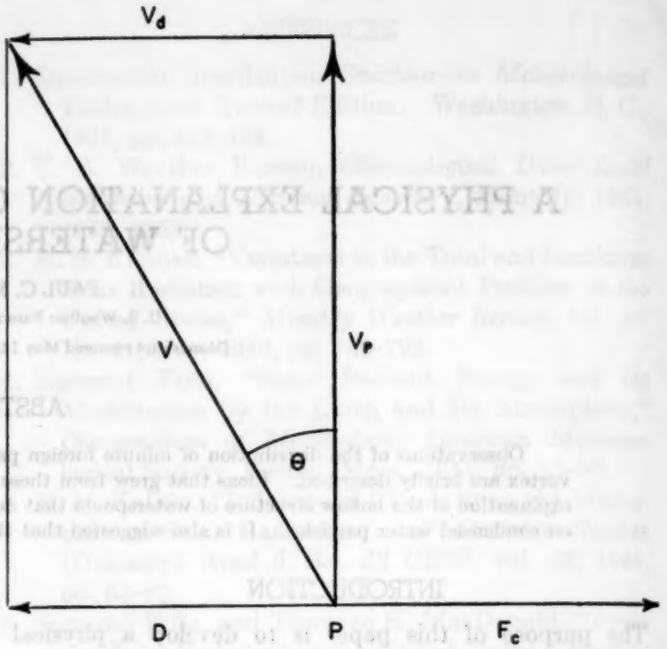

FIGURE 2.—Force and velocity diagram for a spherical particle, P, traveling at constant speed in a circular path (velocity V_p) around a vortex axis at an equilibrium distance where the inward force due to the frictional drag D on the particle caused by a converging flow of air at velocity V is balanced by the outward force on the particle due to centrifugal force F_c. The velocity of the air relative to the velocity of the particle is V_d.

The value of μ will be taken as 0.181×10^{-3} gm. cm.$^{-1}$ sec.$^{-1}$, the viscosity of dry air at 20° C. (c. g. s. units are used throughout this paper). Substituting $M=\frac{\pi\rho_p a^3}{6}$ (where ρ_p is the density of the particle) and $V_p=V_d\cot\theta$, we get

$$r=\frac{\rho_p a^2 V_d\cot^2\theta}{18\mu}. \tag{1}$$

For Reynolds' numbers above 0.5 the general drag formula [4] will be applied using experimentally determined values for the drag coefficient

$$D=\frac{cA\rho V_d^2}{2}$$

c = drag coefficient.
A = projected area of the body in the direction of its motion relative to the fluid.
ρ = density of the fluid (for dry air at 1,000 mb. pressure, 20° C. temperature this equals 1.190×10^{-3} gm. cm^{-3}).
V_d = velocity of the body relative to the fluid.

Again, $D=F_c$ or

$$\frac{cA\rho V_d^2}{2}=\frac{MV_p^2}{r}$$

from which

$$r=\tfrac{4}{3}\frac{\rho_p a}{\rho c}\cot^2\theta \tag{2}$$

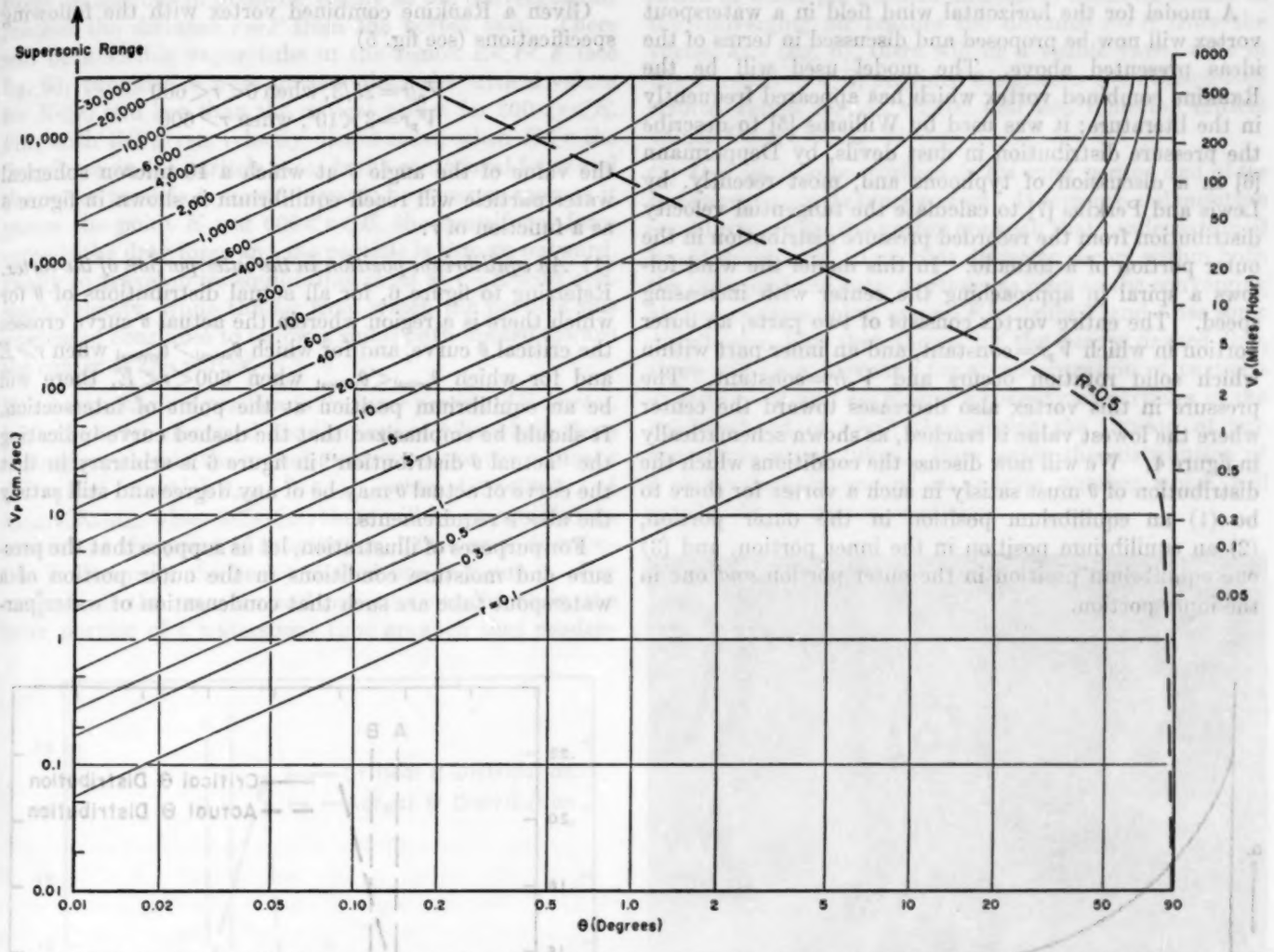

FIGURE 3.—Graph showing the relationship between V_p, r, and θ for water particle 10 microns in diameter. Stokes' law was used for Reynolds' numbers $R \lesssim 0.5$ (for which values of constant r define straight lines). For $R > 0.5$ experimental values for the drag coefficient (see text) were used.

Using relations (1) and (2) it is possible to construct a graph of V_p against θ and draw lines of constant r for spherical particles of a given diameter and density. Figure 3 is such a graph for spherical water particles 10 microns in diameter (about the average size of stratocumulus cloud particles). Formula (1) yielded the straight lines to the left of the line indicating values of Reynolds' number R equal to 0.5; formula (2) gave the curved lines to the upper right. The values of r are plotted in centimeters, hence division by 30 converts them roughly into feet.

As an example of the use of the chart, let us suppose that the particles forming the inside of the visible tube in the laboratory vortex are 10 microns in diameter and that the tube is 1 centimeter in inside radius. Then if the tangential velocity of the air at that part of the vortex is 10 miles per hour, the angle θ must have a value of 7½ degrees. The slope of the lines of constant r on the chart for constant particle velocity indicates that the value of the angle θ decreases as the radius of the hollow tube increases. For example, to produce a hollow tube of radius greater than about 3 feet would require that the angle θ be less than about 3 degrees or V_p supersonic.

WATERSPOUTS

Photographs of waterspouts indicate that hollow tubes are usually observed. Hence, if we accept the above explanation of their occurrence, any model for such a waterspout vortex must provide for convergence in the wind field in the neighborhood of the hollow tube. As mentioned above, figure 3 indicates that there is probably less convergence within the hollow tubes of waterspouts than within the tube in the laboratory model. Another important difference is that the laboratory tube is composed of steam particles introduced into the vortex from below by the pan of steaming water, while the waterspout tube is probably composed of water particles formed near the center of the vortex by condensation due to low pressure in this region.

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A model for the horizontal wind field in a waterspout vortex will now be proposed and discussed in terms of the ideas presented above. The model used will be the Rankine combined vortex which has appeared frequently in the literature; it was used by Williams [5] to describe the pressure distribution in dust devils, by Deppermann [6] in a discussion of typhoons and, most recently, by Lewis and Perkins [7] to calculate the tangential velocity distribution from the recorded pressure distribution in the outer portion of a tornado. In this model the wind follows a spiral in approaching the center with increasing speed. The entire vortex consists of two parts, an outer portion in which $V_p r$=constant, and an inner part within which solid rotation occurs and V_p/r=constant. The pressure in this vortex also decreases toward the center where the lowest value is reached, as shown schematically in figure 4. We will now discuss the conditions which the distribution of θ must satisfy in such a vortex for there to be (1) an equilibrium position in the outer portion, (2) an equilibrium position in the inner portion, and (3) one equilibrium position in the outer portion *and* one in the inner portion.

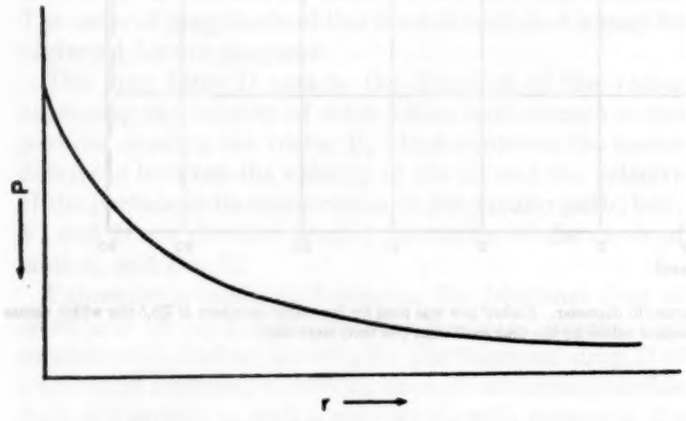

FIGURE 4.—Schematic representation of the horizontal distribution of pressure in the Rankine combined vortex discussed in the text.

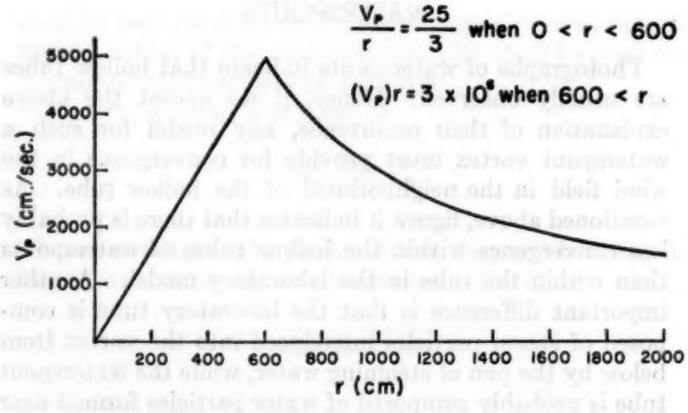

FIGURE 5.—Graph showing the horizontal distribution of tangential velocity for the Rankine combined vortex described in the text (c. g. s. units).

Given a Rankine combined vortex with the following specifications (see fig. 5)

$$V_p/r = 25/3, \text{ when } 0 < r < 600$$
$$V_p r = 3 \times 10^6, \text{ when } r > 600$$

the value of the angle θ at which a 10 micron spherical water particle will reach equilibrium is shown in figure 6 as a function of r.

(1) *An equilibrium position in the outer portion of the vortex.* Referring to figure 6, for all actual distributions of θ for which there is a region wherein the actual θ curve crosses the critical θ curve, and for which $\theta_{actual} > \theta_{critical}$ when $r > E$ and for which $\theta_{actual} < \theta_{critical}$ when $600 < r < E$, there will be an equilibrium position at the point of intersection. It should be emphasized that the dashed curve indicating the "actual θ distribution" in figure 6 is arbitrary in that the curve of actual θ may be of any degree and still satisfy the above requirements.

For purposes of illustration, let us suppose that the pressure and moisture conditions in the outer portion of a waterspout tube are such that condensation of water par-

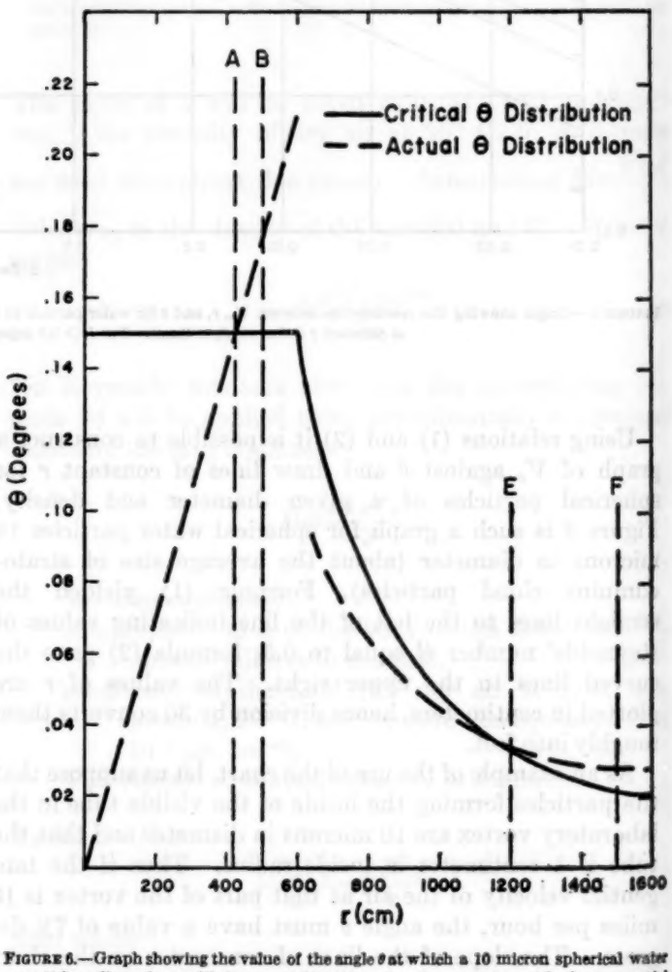

FIGURE 6.—Graph showing the value of the angle θ at which a 10 micron spherical water particle will reach equilibrium as a function of r for a vortex with the tangential velocity distribution shown in figure 5.

ticles 10 microns in diameter occurs as the converging air reaches the distance $r=F$ from the vortex axis. There will be a visible vapor tube in the region $E<r<F$ (see fig. 6), provided θ were greater than the critical values for $E<r$, and less than the critical values for $600<r<E$. For, with the given velocity distribution, when $E<r$ the drag force on the particle exceeds the centrifugal force and the water particle is forced inward; however as soon as it passes the point E and $600<r<E$, the centrifugal force exceeds the drag force and the particle is thrown outward, finally to travel in its equilibrium orbit at $r=E$ with constant velocity, while the air from which the particle was condensed continues to follow a spiral path inward toward the center of the vortex.

(2) *An equilibrium position in the inner portion of the vortex.* In figure 6, for all actual distributions of θ for which there is a region wherein the actual θ curve crosses the critical θ curve, and for which $\theta_{actual}>\theta_{critical}$ when $A<r<600$, and $\theta_{actual}<\theta_{critical}$ when $0<r<A$, there will be an equilibrium position at the point of intersection. One such distribution is shown by the heavy dashed line in figure 6. For example, if the pressure and moisture conditions in the inner portion of a waterspout tube are such that condensation of water particles 10 microns in diameter occurs as the converging air reaches the distance $r=B$ from the vortex axis (fig. 6), there will be a visible vapor tube in the region $A<r<B$, provided θ is greater than the plotted critical values for $A<r<600$ and less than the critical values for $0<r<A$.

(3) *One equilibrium position in the outer portion and at the same time, one in the inner portion.* If pressure and moisture conditions within the vortex are such that condensation of liquid droplets of 10 microns diameter occurs within both the inner and the outer portions as soon as the converging air passes some point, say $r=F$ in figure 7, and θ has some distribution like that shown by the heavy dashed line, there will be a tendency for a concentration of particles at radii $r=A$ and $r=E$, and a sparsity of particles in the neighborhood of $r=C$, thus giving the illusion of two concentric vapor tubes. Here again the distribution of θ shown is only one of an infinite number which will satisfy the necessary conditions.

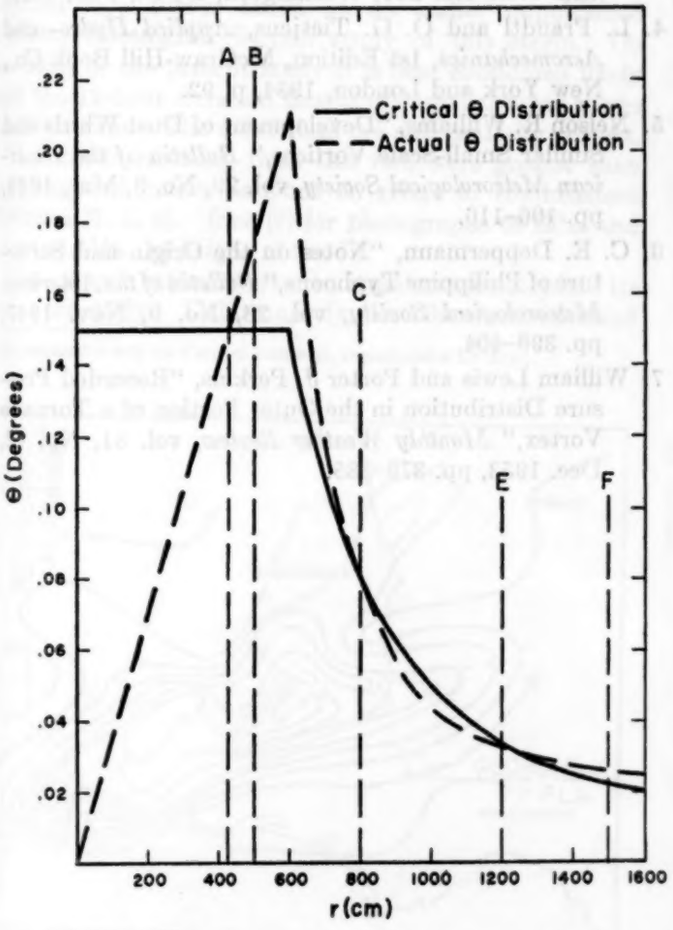

FIGURE 7.—Graph showing a possible distribution of θ which would produce one hollow tube in the inner portion and one hollow tube in the outer portion of a vortex with the tangential velocity distribution shown in figure 5.

FIGURE 8.—Waterspout on the Choptank River Md., 12:20 p. m., August 28, 1938.

Inasmuch as the pressure within the inner portion of our vortex model is lower, condensation would be expected to occur there first during formation; conversely evaporation would be expected to occur first in the outer portion during dissipation. The waterspout pictured in figure 8 occurred on the Choptank River, Md., at 12:20 p. m. on August 28, 1938. The photographer submitted the following remarks with the picture: "Spout beginning to deteriorate. Notice how outside layer is already starting to retreat upward. Spout then about 10 minutes old." It should be emphasized that all of the previous remarks refer only to flow in the vortex above the friction layer. Flow near the surface in waterspout and tornado vortices is probably more much convergent, due primarily to the tendency for particles of air retarded by surface friction to conserve their angular momentum by moving toward the center of rotation.

TORNADOES

It seems reasonable to believe that the mechanism just described also operates in tornado vortices, but that the strength of these vortices is considerably greater than that of waterspouts, so that condensation in the inflowing air occurs far enough from the axis to obscure the hollow structure of the tubes. It is possible that the outside of the tornado funnel marks the outer distance from the axis at which condensation is taking place in the converging air under the existing pressure and moisture distribution in the vortex. The fact that this condensation takes place farther from the axis at higher elevations, thereby producing the characteristic funnel-shaped cloud of the tornado, may be due to one or more of the following effects: (1) higher relative humidities aloft, (2) greater vortex strength aloft, (3) the fact that slightly less lifting is required to saturate air with a given relative humidity at higher elevations.

CONCLUSIONS

The mechanism proposed above to explain the hollow tubes of waterspouts suggests that horizontal convergence is occurring in the waterspout vortex. Inasmuch as there is evidence of the operation of the proposed mechanism in dust devils and tornadoes, it is indicated that horizontal convergence is also a feature of their vortices. Furthermore, if the theory presented here is correct, it suggests that the vertical motion in the visible portion of a waterspout or tornado is upward, not downward. It was found that a converging Rankine combined vortex could be used to explain some observed characteristics of waterspout tubes, however it should be emphasized that converging vortices of other specifications would have served equally well in that part of the discussion.

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THE GREAT-VOLUME RAINSTORM AT ELBA, ALABAMA

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INTRODUCTION

The depth of rain which fell in the Elba, Ala., storm of March 14, 1929, is the greatest in the United States for most durations from 18 to 48 hours over large areas (from 10,000 to 100,000 square miles) [1]. At Elba alone a rainfall total of 21.4 inches was measured in a period of 32 hours ending at 1600 CST, March 14, 1929. Observers' notes and indirect evidence seem to indicate that almost all of this rain fell in the final 12 hours of the period. Figure 1 is the isohyetal map for the 24-hour period starting at 0300 CST, March 14. Figure 2 shows the depth-duration-area curves based on an earlier unpublished Hydrometeorological Section study of this storm. On the basis of the present study it is now considered likely that the 12-hour duration line in the smaller areas shows too little rain. The other lines are substantially correct.

The floods resulting from this rain were greater than any recorded before or since on rivers in southeastern Alabama [2, 3, 4]. (See [2] for photographs of Elba and Brewton, Ala., at height of flood.)

The storm's intensity over so large an area dictated its choice for study. Unfortunately, the Elba storm occurred before the present extensive network of upper-air stations, and so most of our knowledge must come from surface observations. There have been somewhat similar storms (synoptically) since the inauguration of widespread upper-air observations, but of lesser intensity (e. g., the storm centered in Louisiana on April 29, 1953).

A study of the surface maps for the storm period reveals a number of unusual features associated with the occurrence of the phenomenal rainfall. The primary purpose of this paper is to describe these unusual features. Some effort will also be made to offer tentative suggestions concerning the mechanisms involved.

[1] In cooperation with the Corps of Engineers, Department of the Army.

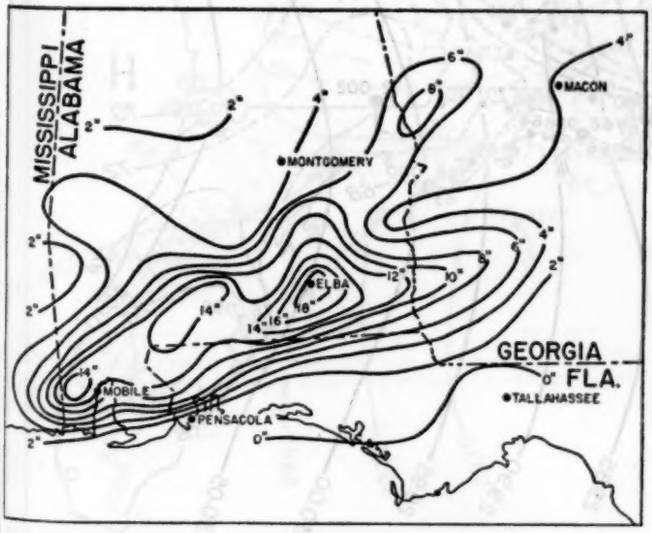

FIGURE 1.—Isohyetal map showin 24-hour rainfall for period starting 0300 CST, March 14, 1929.

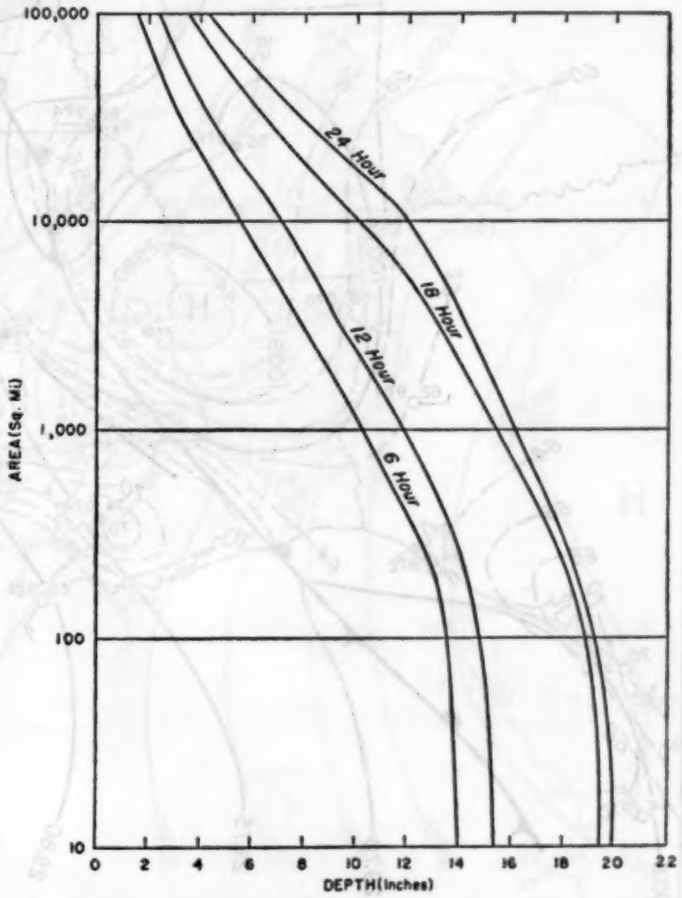

FIGURE 2.—Depth-duration-area curves for the Elba, Ala., rainstorm, March 14, 1929.

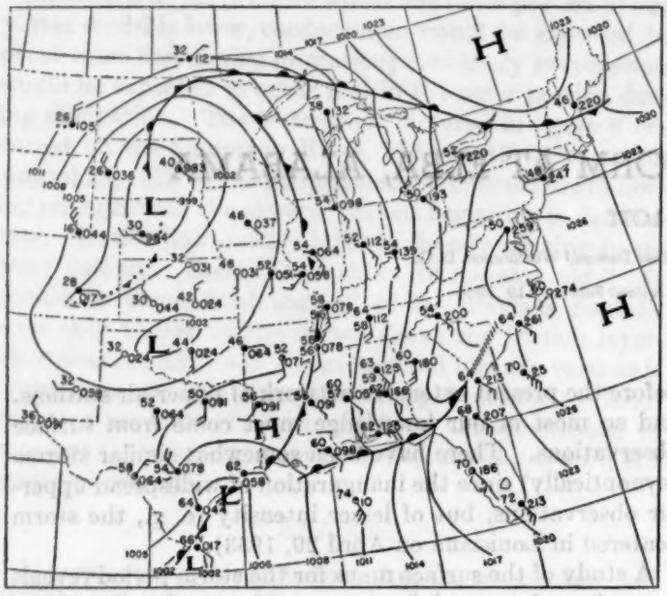

FIGURE 3.—Surface map for 0700 CST, March 13, 1929, about 24 hours prior to the storm at Elba.

GENERAL SYNOPTIC SITUATION

On March 11, 1929, a very deep trough including a cold front extended from about Bismarck, N. Dak., to Phoenix, Ariz., with mP air behind it. To the east, a large High of Arctic origin was centered near Cape Hatteras, directing a broad southerly current over the central part of the country. By March 12, a warm front had appeared in the Gulf of Mexico, with mT air to the south; rain was falling over a large area north of the front. Meanwhile, the deep mid-continent trough had moved slowly eastward with a well-organized surface Low near North Platte, Nebr., and the eastern High had moved slowly off the coast.

Figure 3, the surface weather map about 24 hours prior to the great storm, shows the warm front approaching Pensacola, Fla., from the Gulf. During the afternoon of the 13th the dewpoint at Pensacola rose from 61° to 66° F., while the surface wind shifted from east to south-

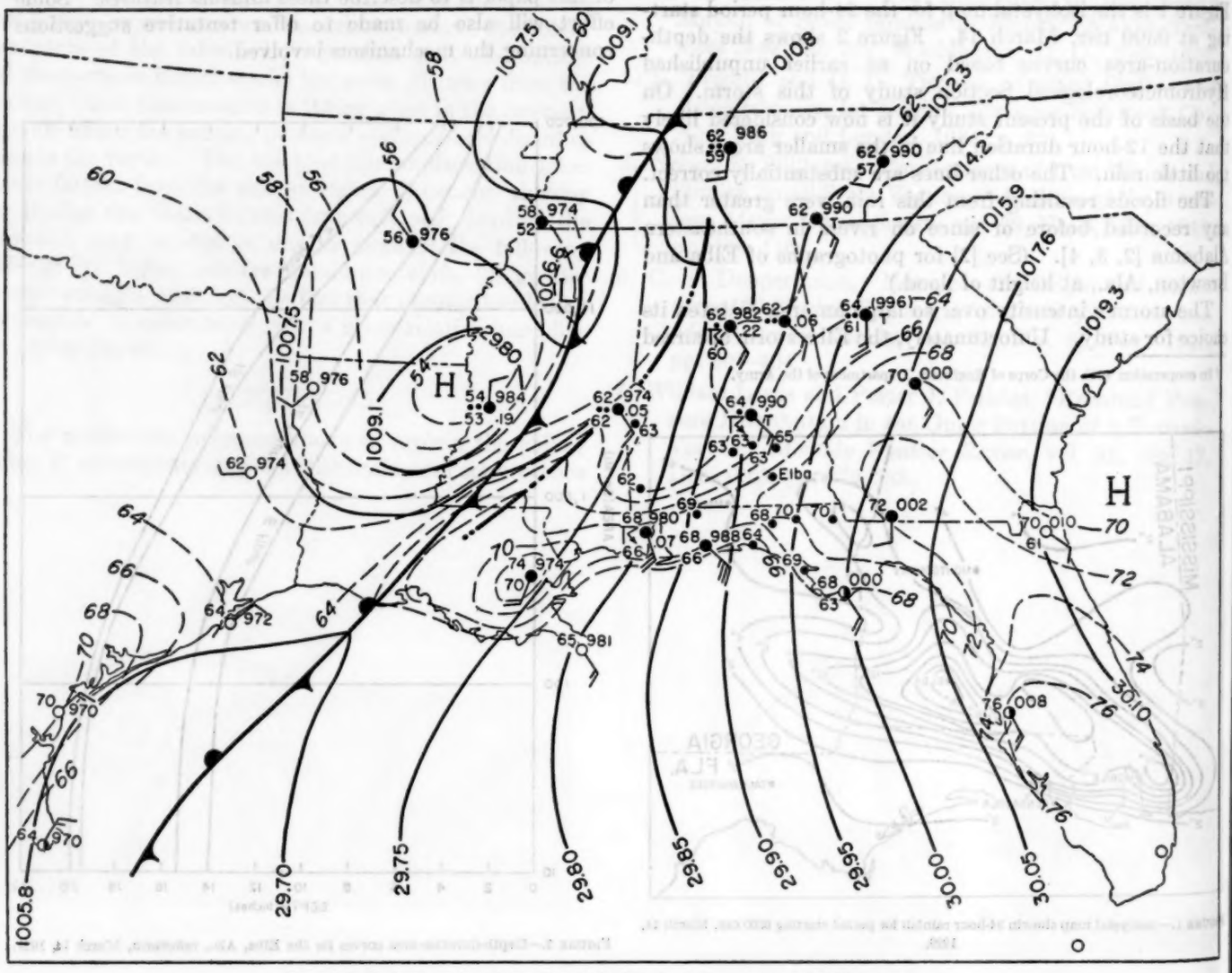

FIGURE 4.—Surface map for 1900 CST, March 13, 1929, about 8 hours prior to the Elba rainstorm.

east. After the frontal passage at Pensacola, no definite position for the warm front could be ascertained. The cold front, which had moved somewhat ahead of the surface mid-continent trough, was oriented north-south near the Mississippi River and extended into the extreme northwestern Gulf. North of Vicksburg, Miss., this front was moving slowly eastward, but in the Gulf of Mexico a very weak stable wave had formed about 50 miles east-northeast of Corpus Christi, Tex. The subsequent history of this stable wave was of great importance in the chain of events that led to the major rain-burst in the Elba storm.

DETAILS OF THE SITUATION

A series of hourly surface weather maps was analyzed for the 24-hour period starting at 1800 CST, March 13, 1929. The maps cover the States of Alabama, Mississippi, Georgia, and northern Florida in as great detail as possible. Figures 4 and 5 represent two maps from this series, those for 1900 CST, March 13 and 0700 CST, March 14. Cooperative observer synoptic weather data, including current temperature readings, for southern Alabama and northwestern Florida were also plotted on the charts. The rainfall amount, where plotted, is for the hour ending at the time of the map. Rainfall intensity is not differentiated by the weather symbol.

Figure 4, illustrating conditions about 8 hours before the deluge, shows a number of interesting features. The stable wave which had formed near Corpus Christi was centered in northern Mississippi, having averaged about 40 m. p. h. in its sprint from the Gulf. Four to five hours after the time of the map the wave passed Nashville, Tenn., causing a maximum wind of 26 m. p. h. from the southeast before the cold-front passage. A shower area, probably induced by the fast-moving wave, formed over Mississippi during late afternoon of the 13th. The line near Meridian, Miss., shows the approximate eastern boundary of this shower area (the instability line). At about 1930 CST this line passed Meridian, and in the next 14 hours passed, successively, Birmingham, Anniston, and Montgomery, Ala., and Atlanta and Macon, Ga. Every

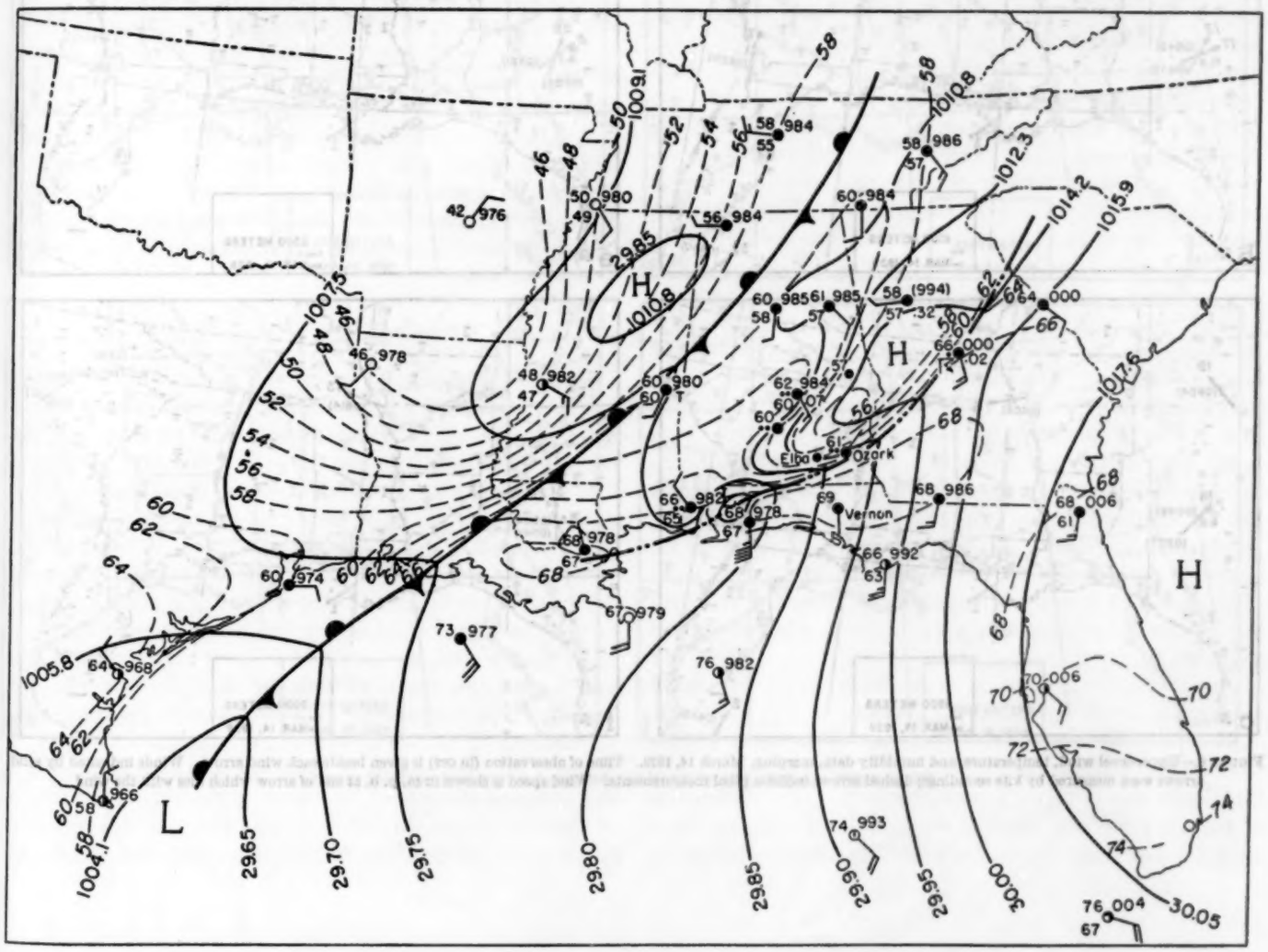

FIGURE 5.—Surface map for 0700 CST, March 14, 1929, the synoptic situation during the heavy rain.

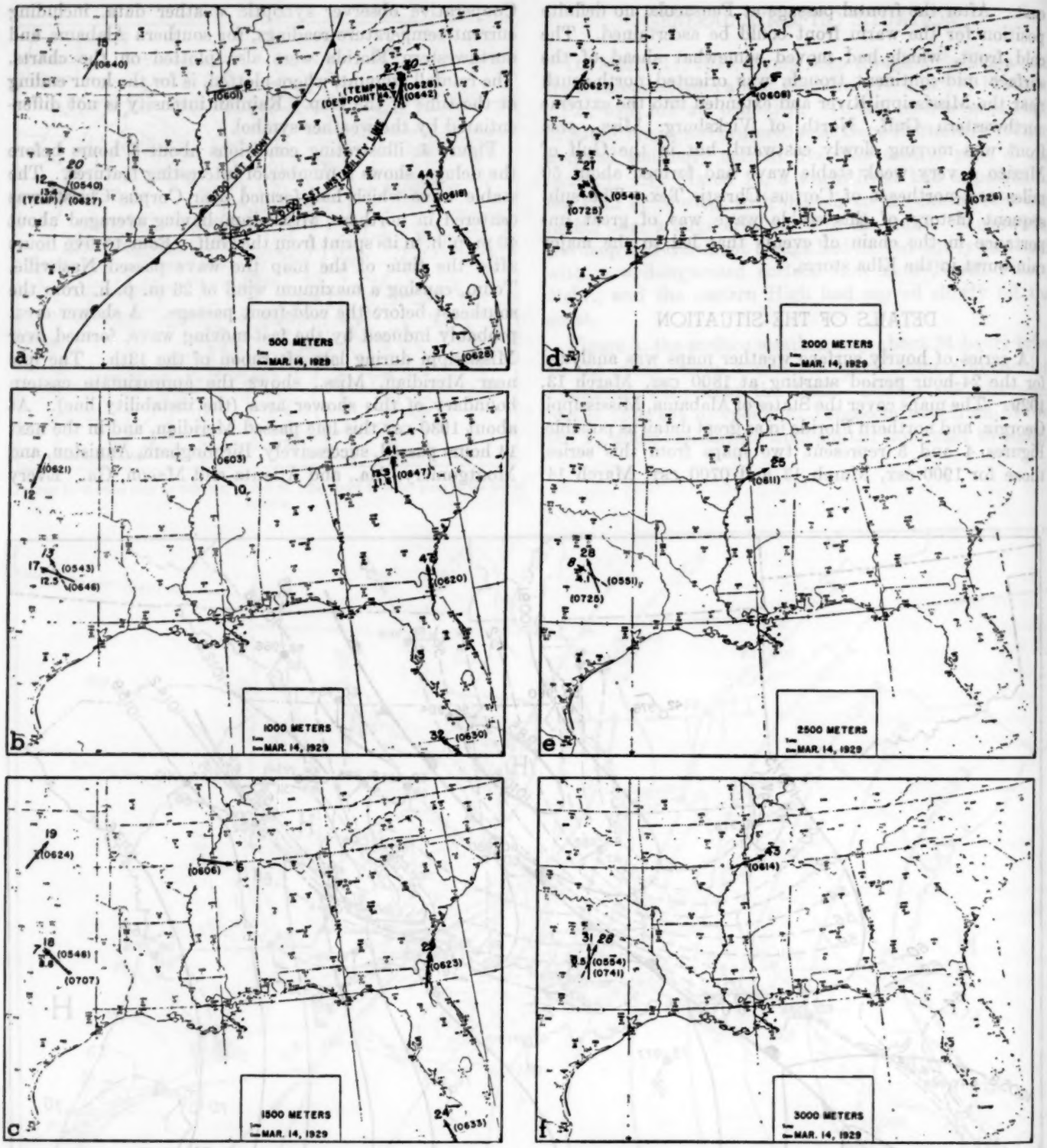

FIGURE 6.—Upper-level wind, temperature, and humidity data, morning, March 14, 1929. Time of observation (in CST) is given beside each wind arrow. Winds indicated by solid arrows were measured by kite soundings; dashed arrows indicate pibal measurements. Wind speed is shown in m. p. h. at end of arrow which flies with the wind.

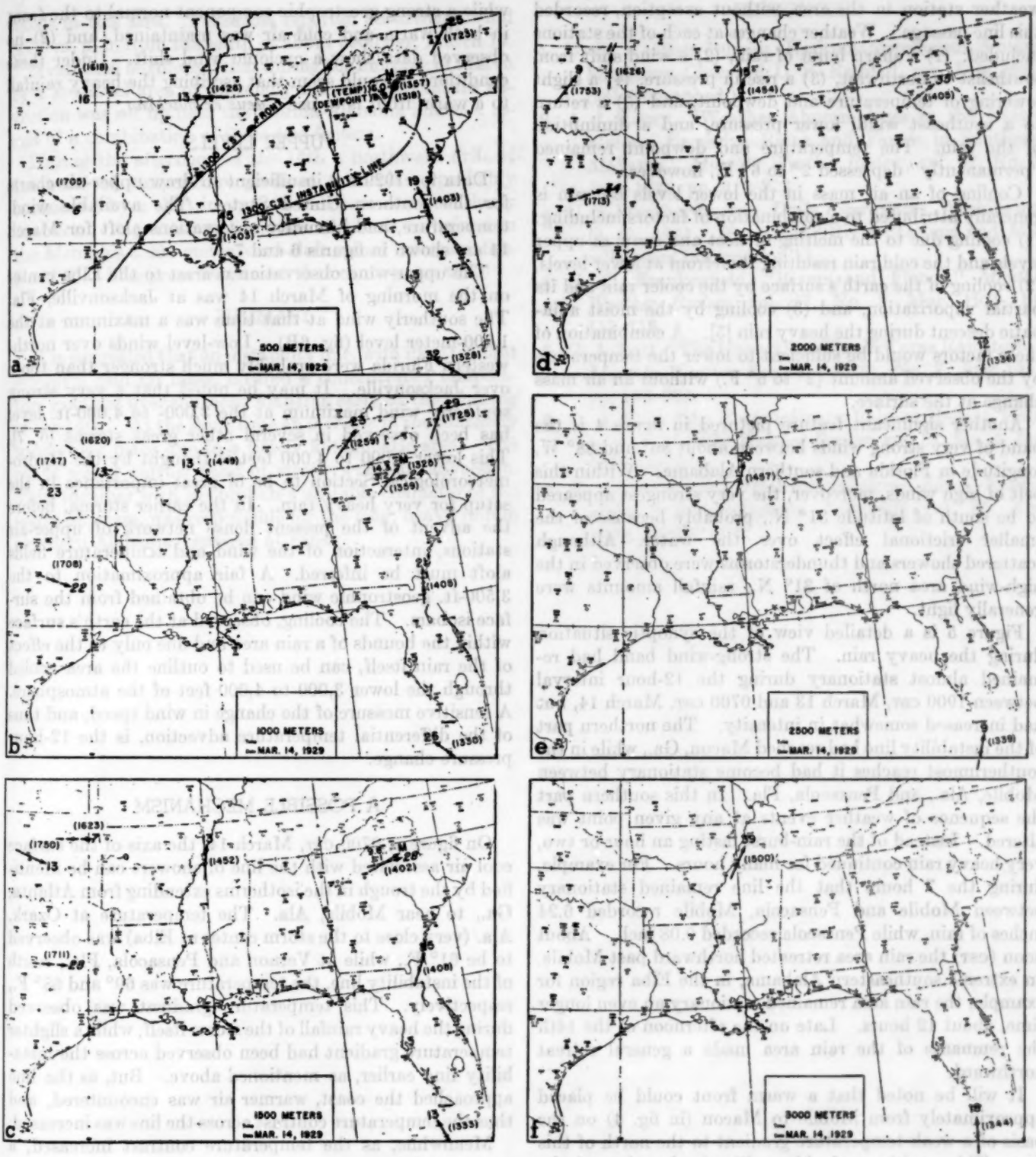

FIGURE 7.—Upper-level wind, temperature, and humidity data, afternoon, March 14, 1929.

weather station in the area, without exception, recorded this line passage. Weather changes at each of the stations included: (1) a sharp burst of rain, (2) a wind shift from southeast to northwest, (3) a rise in pressure, (4) a slight lowering of temperature and dewpoint, and (5) a return to a southeast wind, lower pressure, and a diminution of the rain. The temperature and dewpoint remained "permanently" depressed 2° to 6° F., however.

Cooling of an air mass in the lower levels by rain is generally attributed to a combination of factors including: (1) cooling due to the melting of sleet and snow at upper levels and the cold rain resulting therefrom at lower levels, (2) cooling of the earth's surface by the cooler rain and its partial vaporization, and (3) cooling by the moist adiabatic descent during the heavy rain [5]. A combination of these factors would be sufficient to lower the temperature by the observed amount (2° to 6° F.) without an air mass change at the surface.

Another significant feature pictured in figure 4 is the band of very strong winds between about 86° and 88° W. longitude in Florida and southern Alabama. Within this belt of high winds, moreover, the very strongest appeared to be south of latitude 31° N., probably because of the smaller frictional effect over the water. Although scattered showers and thunderstorms were observed in the high-wind area north of 31° N., rainfall amounts were generally light.

Figure 5 is a detailed view of the synoptic situation during the heavy rain. The strong-wind band had remained almost stationary during the 12-hour interval between 1900 CST, March 13 and 0700 CST, March 14, but had increased somewhat in intensity. The northern part of the instability line had reached Macon, Ga., while in the southernmost reaches it had become stationary between Mobile, Ala., and Pensacola, Fla. In this southern part the sequence of weather events at any given point was altered. Instead of the rain-burst lasting an hour or two, very heavy rain continued for many hours. For example, during the 7 hours that the line remained stationary between Mobile and Pensacola, Mobile recorded 6.24 inches of rain, while Pensacola recorded 0.08 inch. About noon (CST) the rain area retreated northward past Mobile. In extreme southeastern Alabama, in the Elba region for example, the rain area remained stationary an even longer time, about 12 hours. Late on the afternoon of the 14th the remnants of the rain area made a general retreat northward.

It will be noted that a warm front could be placed approximately from Mobile to Macon (in fig. 4) on the basis of a weak temperature gradient to the north of this line. This position coincides with the location of the instability line in figure 5, 12 hours later. The difficulties of placing a classical front in this position for 12 hours are that (1) the front would have had to remain stationary while a strong geostrophic component normal to the front in both warm and cold air was maintained, and (2) no observed data show a cyclonic wind shift. Under these conditions it would seem that ascribing the heavy rainfall to a warm front is to use a *deus ex machina*.

UPPER LEVELS

Data for 1929 are insufficient to draw upper-air charts for the southern United States. The available wind, temperature, and humidity observations aloft for March 14 are shown in figures 6 and 7.

The upper-wind observation nearest to the Elba center on the morning of March 14 was at Jacksonville, Fla. The southerly wind at that time was a maximum at the 1,000-meter level (fig. 6B). Low-level winds over northwestern Florida were probably much stronger than those over Jacksonville. It may be noted that a very strong southerly wind maximum at the 3,000- to 4,000-ft. level has been observed in several other great storms [6, 7]. This lower 3,000 to 4,000 feet is thought by the Hydrometeorological Section to be of great importance in the setup for very heavy rain. In the earlier storms, before the advent of the present dense network of upper-air stations, interaction of the wind and temperature fields aloft must be inferred. A fair approximation to the 3,500-ft. geostrophic wind can be obtained from the surface isobars. The cooling, observed at the earth's surface within the bounds of a rain area and due only to the effect of the rain itself, can be used to outline the area cooled through the lower 3,000 to 4,000 feet of the atmosphere. A sensitive measure of the change in wind speed, and thus of the differential temperature advection, is the 12-hour pressure change.

A POSSIBLE MECHANISM

On figure 5, 0700 CST, March 14, the axis of the surface cool air associated with the line of showers can be identified by the trough in the isotherms extending from Atlanta, Ga., to near Mobile, Ala. The temperature at Ozark, Ala. (very close to the storm center at Elba) was observed to be 61° F., while at Vernon and Pensacola, Fla., south of the instability line, the temperature was 69° and 68° F., respectively. This temperature gradient was observed during the heavy rainfall of the storm itself, while a slighter temperature gradient had been observed across the instability line earlier, as mentioned above. But, as the line approached the coast, warmer air was encountered, and thus the temperature contrast across the line was increased.

Meanwhile, as the temperature contrast increased, a pressure fall area moved in from the west (not shown) and was centered along the Gulf Coast at the longitude of Elba at about 0700 CST, March 14. The effect of this pressure fall was to strengthen further the low-level southerly

winds in this area. Pensacola reported an extreme wind of 60 m. p. h. from the south at about 0900 CST, March 14.

It can then be inferred that during the morning of March 14 a low-level area of unusually intense temperature advection was set up near the Florida-Alabama line as a result of a combination of these two factors.

During the afternoon of the 14th a northward drift of the band of temperature gradient took place, probably occasioned by a shift in upper winds to a more southerly direction. A partial confirmation of this is contained in the Memphis 3,000-meter wind-aloft record (compare figs. 6 and 7). The great rain-burst thus came to an end.

SUMMARY

It is suggested that an instability line, with its attendant low-level cold air mass, created a concentrated temperature gradient over southern Alabama early on the morning of the 14th. At the same time an increase in the southerly winds took place due to a timely katallobaric area. The result was a narrow band of strong warm-air advection. This, together with the probably unstable character of the warm air mass, was responsible for the long, heavy rain in southern Alabama, centered at Elba.

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A PREFERRED THICKNESS LINE ACCOMPANYING MULTIPLE TORNADO OCCURRENCES

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The purpose of this note is to call attention to a preferred 1,000–500-mb. thickness line that in the mean accompanies multiple tornado outbreaks in the United States.

Studies by Sutcliffe [1] and others have suggested that thickness patterns are a suitable synoptic tool for obtaining a picture of the three-dimensional structure of the atmosphere. Sutcliffe and Forsdyke [2] have placed particular emphasis on charts showing the pattern of thickness of the 1,000–500-mb. layer. The contribution of the thickness pattern and the synoptic pressure patterns to the vorticity of the tornado is outside the scope of the present study. However, though much has been written concerning the value of such patterns in the evaluation of vertical motion and synoptic development, there has been little mention of the possible forecasting significance of the position of particular thickness lines, which are equivalent to the mean temperature of the air columns considered.

That a preferred thickness line should accompany tornado outbreaks may be inferred from the findings of several investigators and is a vital part in a possible mechanism for the generation of violent vertical convection that is here postulated. This mechanism is pictured as an air column of given mean temperature in the troposphere being restrained to that mean temperature while at the same time warm air is being injected into low levels; this produces an air column which is extremely unstable in the vertical. Unpublished studies, made by the U. S. Weather Bureau at Memphis, Tenn., under the supervision of Mr. Gilley T. Stephens, have shown that during tornado occurrences in that area such a condition does apparently prevail. At 500 mb. the temperature holds constant between −10° and −16° C. and at 850 mb. there are large positive increases in temperature and humidity. Therefore, since it is known [3] that most tornadoes occur with surface dew points exceeding 55° F., there probably exist various sets of mean air column temperatures such that introduction of such high moisture values at low levels would create vertical instability.

A mechanism for holding the mean temperature, or as we shall see "thickness," pattern constant during such an intrusion of "warmer" air at low levels can be visualized as being of two possible types. First, the thickness lines could be very closely spaced indicating a marked temperature contrast and a mass of cold air near the ground which would offer some frictional resistance to rapid displacement. Secondly, the approach of stronger winds at the 500-mb. level in the direction of the thickness lines would tend to maintain the strong wind shear represented by the thickness lines, thus furnishing a possible mechanism for maintaining, temporarily at least, the "status quo" of the thickness pattern. Some support for the latter view is found in a study by Schmidt [4] of tornado forecasting for western Tennessee and western Kentucky. He has found that one necessary prior condition to the occurrence of tornadoes is an increase in westerly momentum at 500 mb. to the west of the area.

Some further evidence of the possible existence of a set of preferred thickness lines can be deduced from (1) Fletcher's [5] discovery that wind maxima or "jets" at 500 mb. tend to follow preferred contours, and (2) the known connection between tornadoes and relatively low sea-level pressures or 1,000-mb. contours.

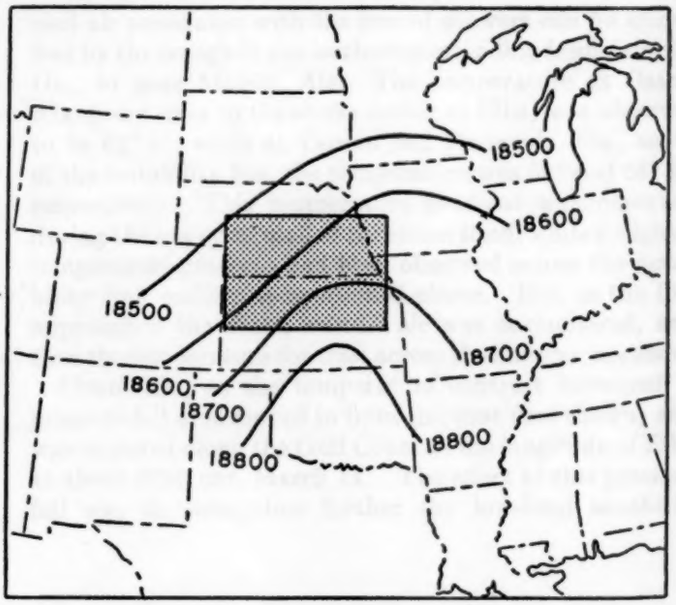

FIGURE 1.—1,000-500-mb. thickness computed from composite charts of 13 tornado situations in the shaded area.

It was therefore decided to make a preliminary determination of the possibility of the existence of such a line, or lines, through a study of composite thickness patterns associated with tornadoes.

The Weather Bureau [6] has prepared charts showing composite contour patterns which accompanied, or preceded by not more than 12 hours, multiple tornado outbreaks in five areas of the Midwest and South.[1] From the composite 1,000-mb. and 500-mb. patterns thus furnished the 1,000–500-mb. thickness patterns were constructed for the five areas shown in figures 1 through 5. Note the consistent presence of the 18,600-ft. thickness line in each area. This line, incidentally, has been found also to be roughly parallel to the normal tracks of tornadoes in the areas shown.

The relationship shown here is, of course, a relationship between the averages of groups of data and not between individual cases. On the basis of the arguments previously given it would appear that there should be combinations of 1,000–500-mb. thickness and low level humidity involving thickness patterns other than those in

[1] A "major tornado day" or multiple tornado outbreak was arbitrarily defined for inclusion in the composite data [5] as one in which three or more tornadoes occurred in one area when at least two of these tornadoes were separated by a distance of 100 miles or more. There were a few days included in these data in which a "major tornado day" occurred in two areas so that data in these cases were included twice. Most of the occurrences during the six tornado seasons (March through June) of 1945 through 1950 that met the requirements for a major tornado day were used.

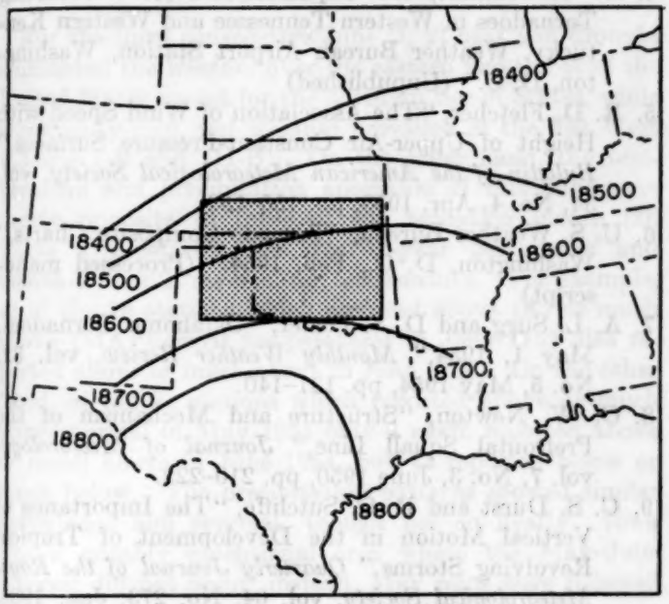

FIGURE 2.—1,000–500-mb. thickness computed from composite charts of 28 tornado situations in the shaded area.

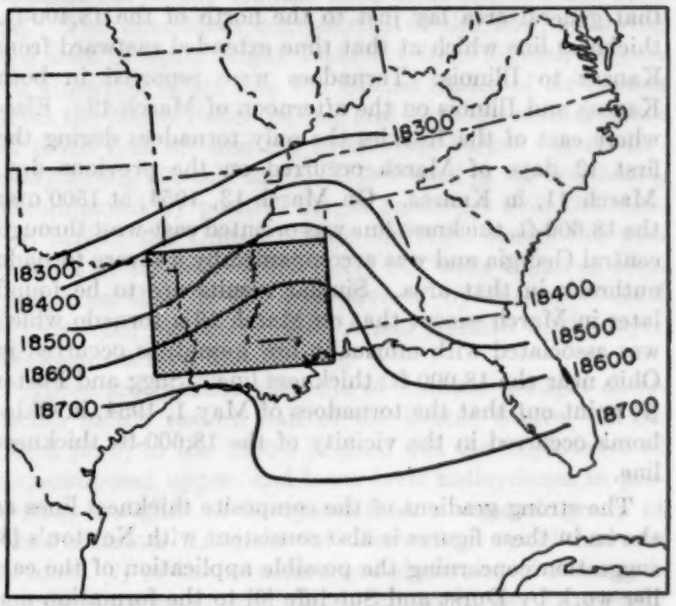

FIGURE 4.—1,000–500-mb. thickness computed from composite charts of 10 tornado situations in the shaded area.

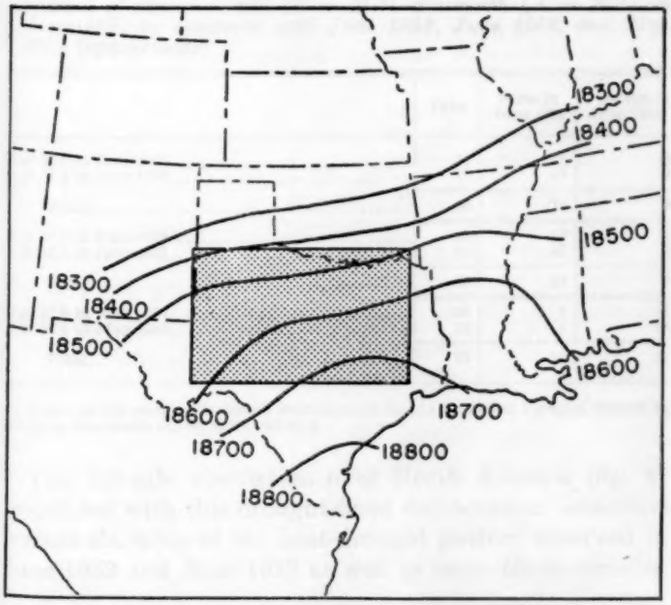

FIGURE 3.—1,000–500 mb. thickness computed from composite charts of 13 tornado situations in the shaded area.

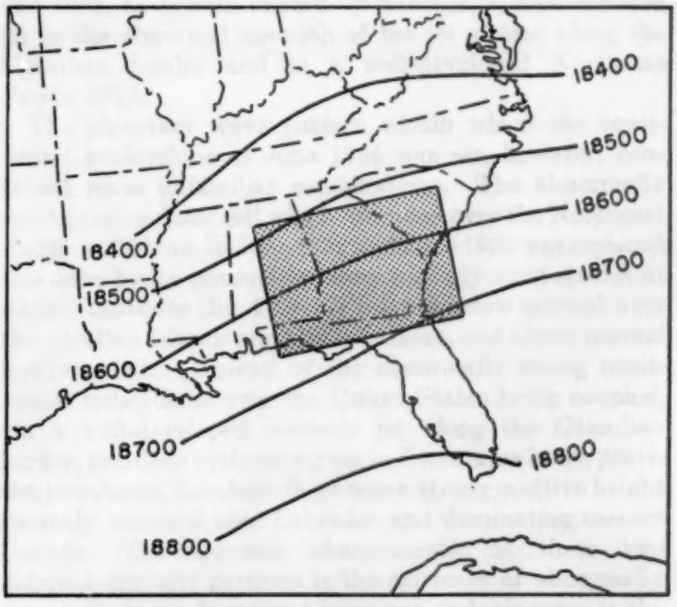

FIGURE 5.—1,000–500-mb. thickness computed from composite charts of 8 tornado situations in the shaded area.

the immediate vicinity of 18,600 ft. which would result in instability. It would also appear that the indicated strong thickness gradient would not always be a necessary accompaniment of the 18,600-ft. line. Examination of daily thickness charts and tornado data for March 1954 shows this to be true while at the same time continuing to point to the special significance of the 18,600-ft. line. During the first part of March 1954 this line remained south of the United States east of the Rockies until 1500 GMT of the 10th when it appeared in the Texas-Oklahoma Panhandle area. By 0300 GMT of the 12th it had reached northward just beyond Oklahoma City and by 1500 GMT of the 12th it was located just north of Memphis, Tenn. However at that time the strongest thickness gradient in that general area lay just to the north of the 18,400-ft. thickness line which at that time extended eastward from Kansas to Illinois. Tornadoes were reported in both Kansas and Illinois on the afternoon of March 12. Elsewhere east of the Rockies the only tornadoes during the first 12 days of March occurred on the previous day, March 11, in Kansas. On March 13, 1954, at 1500 GMT the 18,600-ft. thickness line was oriented east-west through central Georgia and was accompanied by a severe tornado outbreak in that area. Similar results are to be found later in March except that on March 19 a tornado which was associated with unusually low humidites occurred in Ohio near the 18,000-ft. thickness line. Sugg and Foster [7] point out that the tornadoes of May 1, 1954, in Oklahoma occurred in the vicinity of the 18,600-ft. thickness line.

The strong gradient of the composite thickness lines as shown in these figures is also consistent with Newton's [8] suggestion concerning the possible application of the earlier work by Durst and Sutcliffe [9] to the formation and maintenance of squall lines.

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THE WEATHER AND CIRCULATION OF JUNE 1954[1]

Illustrating the Birth and Growth of a Continental Anticyclone

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THE MONTH AS A WHOLE

For the third consecutive June [1, 2] heat and drought dominated the weather over the eastern two-thirds of the United States except for the northern tier of States, while abnormally cool, wet weather prevailed in the Far West (Charts I-B and III-B). On the other hand, the temperature and precipitation anomalies of this June were nearly opposite to those of the preceding month [3]. Table 1 illustrates this year-to-year persistence and month-to-month reversal for temperature. For example, 43 of the 76 stations which reported above (A) or much above (MA) normal temperatures in June 1952 also reported above or much above in June 1954. On the other hand, of the 21 stations which reported above or much above normal in May 1954 only 4 again reported above or much above in June 1954, but 13 reported below or much below. The precipitation patterns showed similar persistence and reversal, though not so neatly. Iowa was a notably exceptional area where the May–June reversal did not occur; instead, May rains were followed by June rains to produce record floods.

TABLE 1.—*Number of United States stations*[1] *showing repetition or reversal of abnormal cold (B or MB) or warmth (A or MA) in June 1954 as compared with June 1952, June 1953, and May 1954 (approximate)*

	Total	Same in June 1954	Opposite in June 1954
B or MB in June 1952	14	11	0
A or MA in June 1952	76	43	9
Total	90	54	9
B or MB in June 1953	21	17	1
A or MA in June 1953	65	41	2
Total	86	58	3
B or MB in May 1954	66	6	41
A or MA in May 1954	21	4	13
Total	87	10	54

[1] Based on 100 evenly distributed stations used by the Extended Forecast Section in verifying prognostic temperature patterns.

The 700-mb. circulation over North America (fig. 1) associated with this drought-flood combination contained certain elements of the heat-drought pattern observed in June 1952 and June 1953 as well as some characteristics reminiscent of the Kansas flood situation of June 1951 [4]. The anomalous components of the flow, following the lines of equal height departure from normal, suggest an abnormal influx into the United States of air masses from the North Pacific Ocean, and less than normal influx of moist Gulf air into the Mississippi Valley. A moderately strong 700-mb. anticyclone existed in the Southeast, accompanied by positive height anomalies over the eastern half of the United States. A High center was also present at sea level (Chart XI) but with negative pressure anomalies in the area covered by the upper-level anticyclone, indicative of abnormal warmth in the layer between sea level and 700 mb.

These are familiar elements of summer drought circulation in the eastern half of the United States. T. R. Reed [5, 6] in the 1930's pointed out the importance of superimposed upper- and lower-level anticyclones in producing drought. A few years later a research group at Massachusetts Institute of Technology published an account [7] of the structure and dynamics of the continental summer anticyclone which could not be greatly improved upon today. Recent occurrences of this anticyclone, associated with the summer droughts of 1952 and 1953, have been related by Klein [1, 8] and Winston [2] to the abnormal strength of the jet stream along the Canadian border and to a well-developed Northeast Pacific High.

The planetary wave pattern within which the continental anticyclone of June 1954 was set, however, contained some unfamiliar combinations. The abnormally strong anticyclonic cell which extended over the Northeast Pacific at 700 mb. in June 1952 and June 1953 was replaced this June by an unusually strong westerly wind stream at middle latitudes (fig. 2), with heights below normal over the Aleutian Islands and Gulf of Alaska, and above normal farther south. Instead of the abnormally strong continental anticyclone over the United States being coupled, via a well-developed westerly jet along the Canadian border, to a deep cyclonic vortex in Canada, as in the previous two Junes, this June there was a strong positive height anomaly centered near Labrador and dominating eastern Canada. The common characteristic of these two different drought patterns is the existence of abnormally strong upper winds tangent in an anticyclonic sense to the

[1] See Charts I-XV following p. 172 for analyzed climatological data for the month.

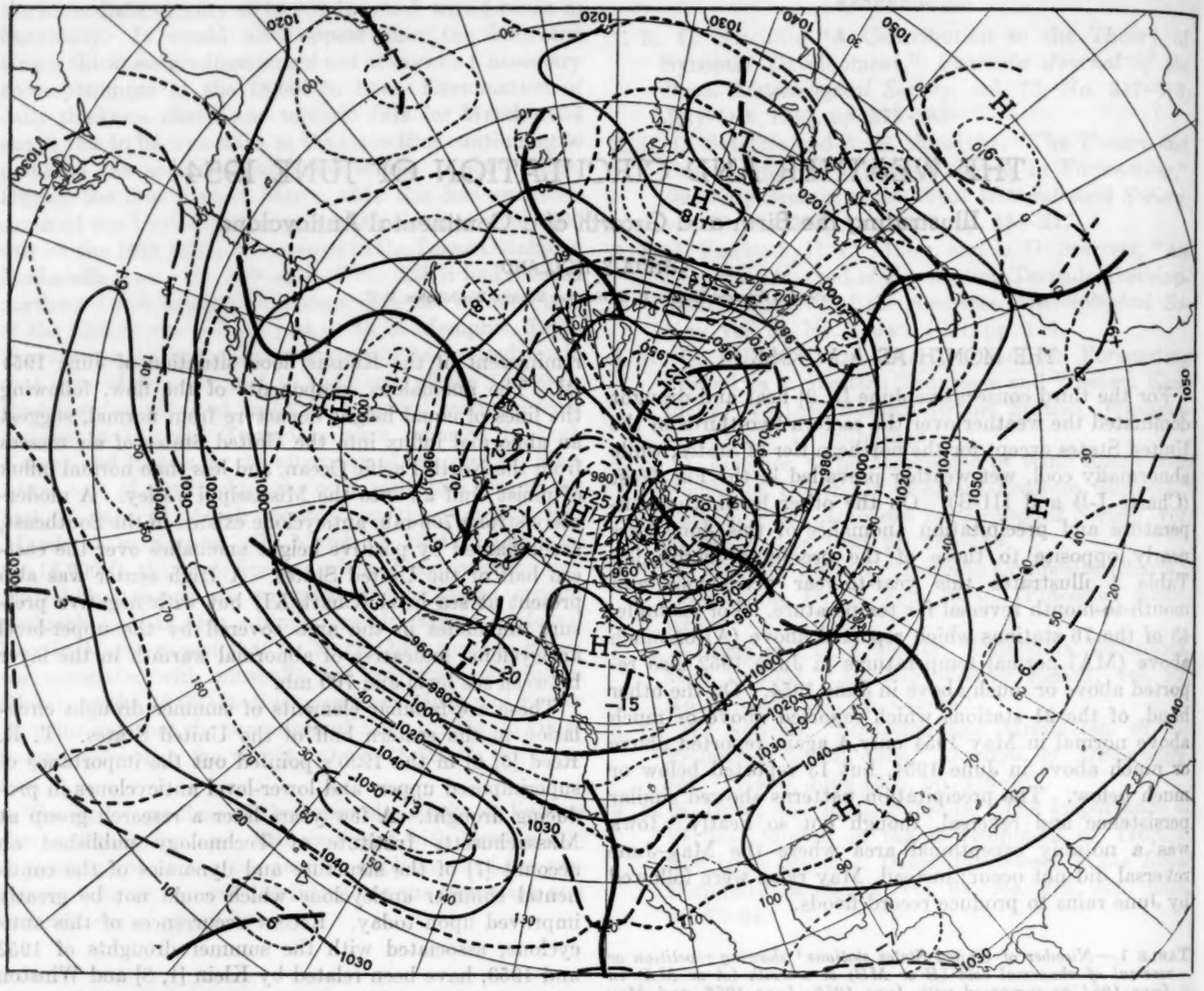

FIGURE 1.—Mean 700-mb. height contours (solid) and departure from normal (dashed) (both in tens of feet) for June 1–30, 1954. Abnormally strong Aleutian Low and East Pacific High poured cold Pacific air into the West while the circulation in the Atlantic and Europe was characterized by blocking, with anomalous cold troughs at low latitudes.

continental seat of the high-level anticyclone. In 1952 and 1953 there were strong westerlies on its north, while in 1954 there were strong southwesterlies on its northwest and northerlies on its east. It is of interest that the June droughts of 1933 and 1936 were accompanied by strongly anomalous southerly components along the west coast. This suggests that the banking mechanism postulated by Rossby [9], by means of which air is piled up to the right of strong, presumably supergradient winds, may be of primary importance in maintaining the continental anticyclone.

The principal features of this month's circulation which persisted from the preceding month were the abnormally fast Pacific westerlies (fig. 2) and the eastern Canadian "blocking High" (260-ft. positive height anomaly in fig. 1). The fast westerlies had shifted eastward, however, to pour cold Pacific air into the western United States, where the weak May circulation had permitted abnormal continentality and therefore warmth. The negative anomaly center in the eastern United States associated with blocking in May had given way to a summer type continental anticyclone, bringing fair, warm weather to the Midwest. The eastern Canadian block, together with a deeper than normal trough off the east coast of the United States, produced recurrent "back-door" cold front passages along the Atlantic seaboard, similar to, but not as strong as, those of June 1953. Thus an eastward shift of the Pacific westerlies and the building up of a continental upper-level anticyclone were the main circulation changes associated with the May-June temperature reversal, carrying on the unusual sequence of month-to-month reversals which began last January [3].

Floods occurred on the Des Moines River in Iowa from June 17 to 24, and on the Rio Grande and its tributaries

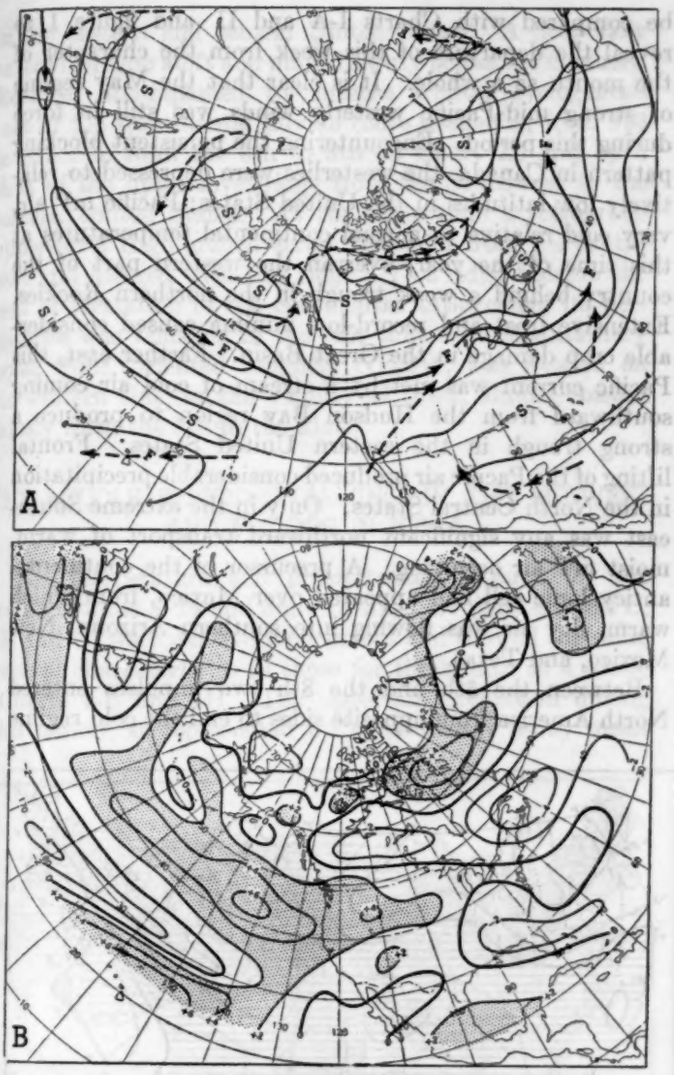

FIGURE 2.—(A) Mean 700-mb. isotachs and (B) departure from normal wind speed (both in meters per second) for June 1-30, 1954. Solid arrows indicate average position of the main 700-mb. jet stream, while dashed arrows indicate secondary axes of relative maximum wind speed. The middle-latitude westerly jet was abnormally strong in the East Pacific but weak in the Atlantic.

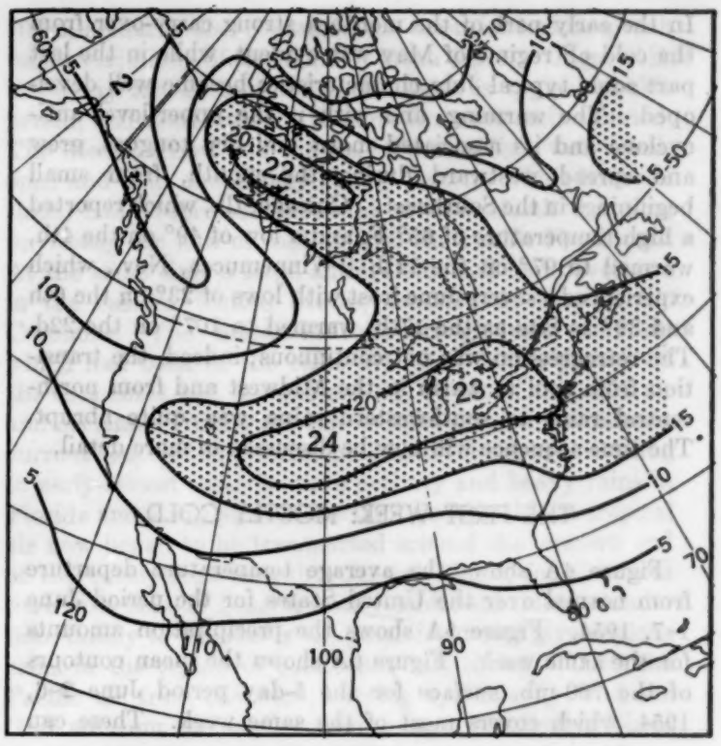

FIGURE 3.—Number of days in June 1954 with surface fronts of any type (within squares with sides approximately 500 miles). Frontal positions taken from *Daily Weather Map*, 1:30 p. m., EST. Heavy rains occurred in the North Central States where fronts were present as much as 80 percent of the time.

in southern Texas from June 27 to 30. In between these areas record doughts occurred, with Fort Smith, Ark., for example, reporting only 0.38 in. of rain for the month, equal to the previous low June reading obtained in 1914. In a gross way these can be associated with certain features of the mean monthly circulation. Iowa is located in the region where the cold Pacific westerlies met the moist southwesterly current flowing around the western end of the continental anticyclone. There was also a zone of maximum frequency of surface fronts running from eastern Colorado across Iowa through the Great Lakes during the month (fig. 3). Chart III-B shows above normal precipitation along this zone from eastern South Dakota across the Lakes. Although somewhat off the axis of this zone of maximum frequency of fronts, Devils Lake, N. Dak., reported a record high June rainfall of 8.55 in. as compared with a previous high of 5.84 in. in 1914.

The Rio Grande floods can be associated with the onshore, cyclonic character of the anomalous component of the monthly mean flow across the west Gulf Coast at both 700 mb. (fig. 1) and sea level (Chart XI inset). These anomalies reflect the brief but strong cyclonic circulation accompanying the passage of Hurricane Alice, superimposed on a relatively normal pattern for the rest of the month. The drought in Oklahoma, Arkansas, and Missouri can be attributed to a peristent anticyclonic flow regime in this area rather than to lack of moisture, since this area was under the influence of both moist and dry air masses at different times.

Along the eastern seaboard, despite frequent frontal activity (fig. 3) record droughts occurred. Norfolk, Va., reported a total of 0.37 in. of precipitation for the month, well below the previous low of 1.05 in. recorded in 1921, and Savannah, Ga., reported a new record low of 0.84 in. compared with a previous low of 0.91 in. in 1891. Near-record low amounts were recorded over a wide arc from Washington, D. C., to Pensacola, Fla. This is not surprising when it is realized that little atmospheric moisture was available during most of the month, the flow having been predominantly from northerly directions with anticyclonic curvature.

Despite the accumulation of the extreme anomalies which have been mentioned so far, the weather of this month did not have a uniform character throughout. In fact, only in the Pacific Northwest were temperature departures of the same sign throughout the month.

In the early part of the month a strong carry-over from the cold cP regime of May was evident, while in the last part some typical July characteristics became well developed. The warming, and with it the upper-level anticyclone and its associated moist and dry tongues, grew and spread westward during the month, from small beginnings in the Southeast. Chicago, Ill., which reported a high temperature of 65° F. and a low of 49° on the 4th, warmed to 97° on the 11th. Winnemucca, Nev., which experienced a severe June frost with lows of 23° on the 6th and 39° as late as the 17th, warmed to 107° on the 22d. The retrogression was not continuous; indeed, the transition from cool to warm in the Midwest and from north-central rains to southwestern rains was quite abrupt. The time sequence will now be discussed in more detail.

THE FIRST WEEK: MOSTLY COLD

Figure 4A shows the average temperature departure from normal over the United States for the period June 1–7, 1954. Figure 5A shows the precipitation amounts for the same week. Figure 6A shows the mean contours of the 700-mb. surface for the 5-day period June 2–6, 1954, which covers most of the same week. These can be compared with Charts I-A and II and figure 1 to reveal the departure of this week from the character of the month as a whole. It is clear that the May regime of strong mid-Pacific westerly winds was still in force during this period. Encountering the persistent blocking pattern in Canada, the westerlies were depressed to relatively low latitudes in the United States; Pacific mP air, very cold relative to normal continental temperatures at this time of the year, overran the western part of the country behind a weak trough in the northern Rockies. Extensive frost and record-low minima caused considerable crop damage in the Great Basin. Farther east, this Pacific current was met by a stream of cold air coming southward from the Hudson Bay region to produce a strong trough in the eastern United States. Frontal lifting of the Pacific air produced considerable precipitation in the North Central States. Only in the extreme Southeast was any significant northward transport of warm, moist mT air occurring. A precursor of the continental anticyclonic cell had appeared over Mexico, from which warm, dry air was flowing into southern Arizona, New Mexico, and Texas.

Between the 5th and the 8th two impulses entered North America from opposite sides to end the cold regime

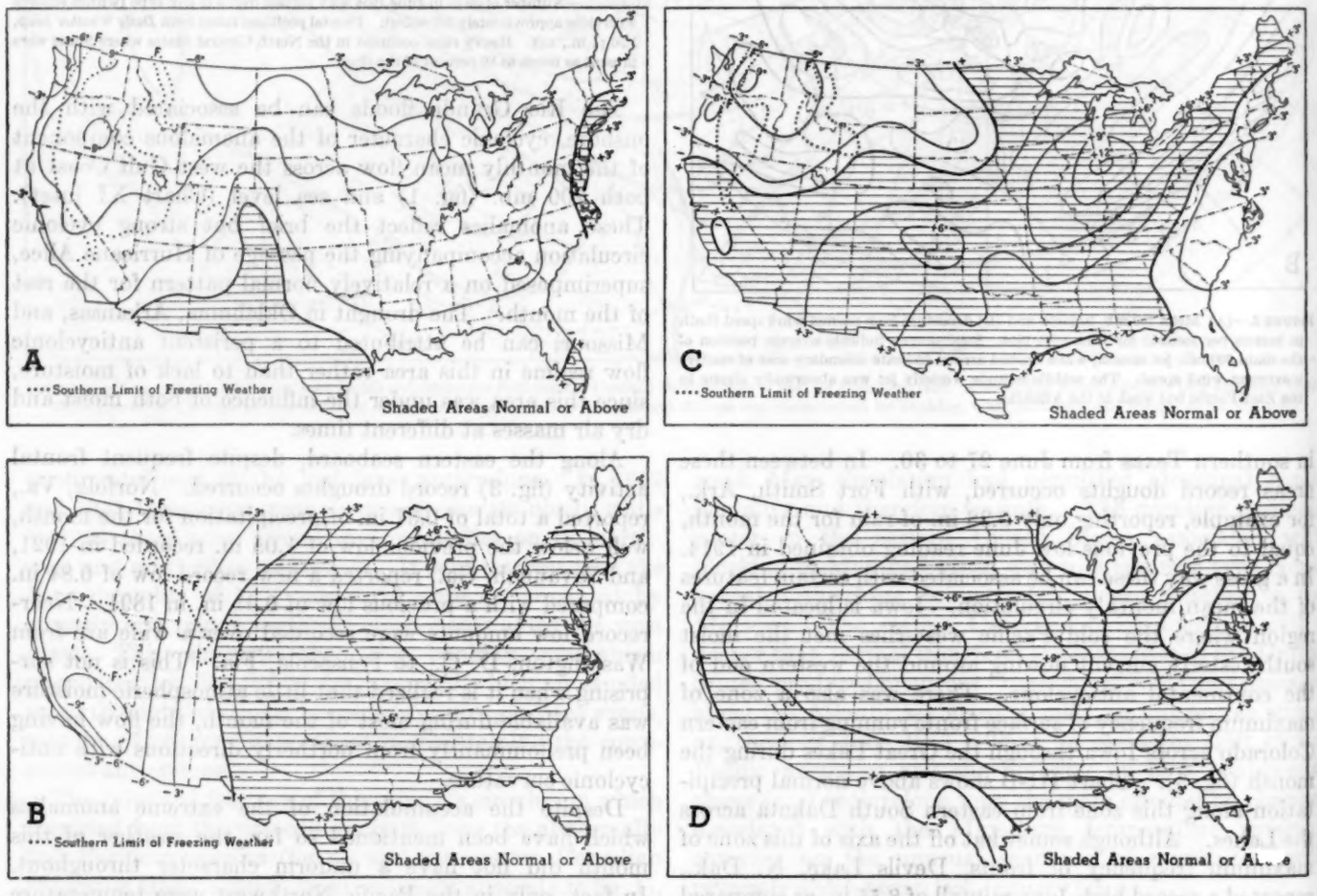

FIGURE 4.—Departure of average temperature (°F.) from normal for the weeks of (A) June 1–7, (B) June 8–14, (C) June 15–21, and (D) June 21–28, 1954. Heat replaced cold as the continental high-level anticyclone spread westward.

in the eastern and central United States. A deep Low entered Oregon aloft and moved slowly across the northern Rockies, accompanied by intense frontal cyclogenesis east of the Continental Divide (note deepening from 993 to 975 mb. from the 6th to 8th on Chart X) and by a shift of the upper-level ridge from the Southwest to the eastern United States. At the same time a retrograde upper-level High arched across Labrador from the southeast and merged with this ridge. This ridge became the seat of formation of the continental anticyclone; its persistence was reflected in the fact that a quasi-stationary polar front, oriented roughly NE-SW, was located on its western side, in the vicinity of Iowa, almost continuously from the 8th to the 21st.

THE MIDDLE PERIOD: WARMING IN THE MIDWEST AND FLOODS IN IOWA

The temperature and precipitation patterns for the weeks of June 8–14 and June 15–21, 1954 (figs. 4B and C, 5B and C) were relatively persistent and showed the same general characteristics as those for the month as a whole (Charts I-B and II). The upper-air circulation during this period, as represented by twice-weekly 5-day mean 700-mb. charts, also contained the essential features, with minor variations, of the monthly mean (fig. 1). Figure 6B, the 5-day mean chart for June 16–20, is typical of this period, and illustrates the changes from the earlier period. The Aleutian Low and the East Pacific High, well developed and with abnormally strong westerly winds between them, had progressed eastward toward North America, increasing the flow of cold air into the Pacific Northwest. At the same time a large anticyclonic cell had developed in the eastern United States, centered at first in the Carolinas. The middle-latitude trough and cold air which had been in the eastern United States passed off the east coast and became cut off. Part of this cold air and cyclonic vorticity appears to have entered the easterly current south of the continental anticyclone, contributing to early-season easterly wave activity and heavy rains in Florida and along parts of the Gulf Coast. Moist tropical air now began to be transported around the western end of the anticyclone into the Central Plains, for the most part in a field of general anticyclonic, subsiding circulation, finally to encounter the polar front and give copious rains in the region of confluence and lifting. Both the Pacific air from the west, curving cyclonically, and the Gulf air from the south, curving anticyclonically, joined

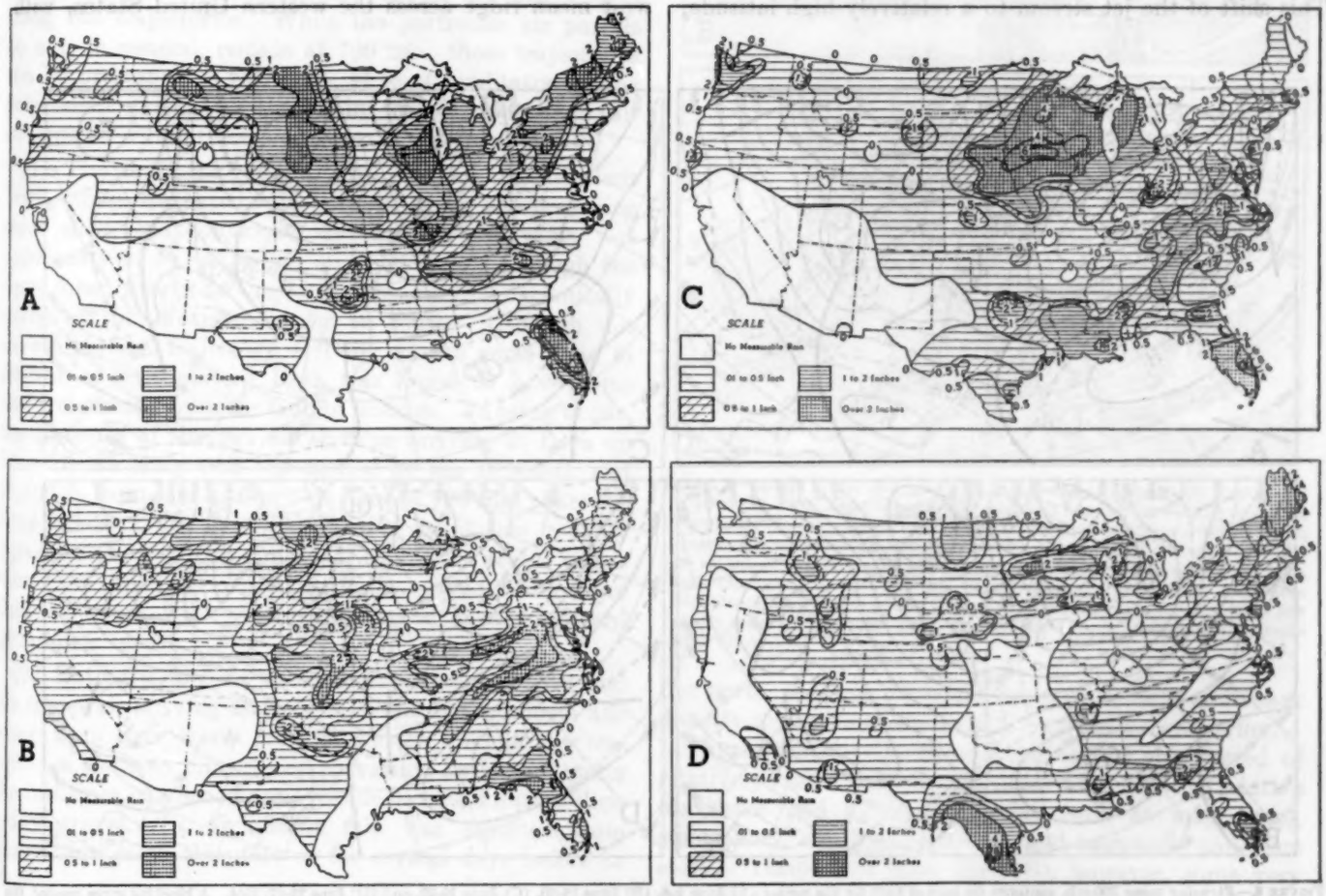

FIGURE 5.—Total precipitation (inches) for the weeks of (A) June 1-7, (B) June 8-14, (C) June 15-21, and (D) June 22-28, 1954. The rains shifted westward as the continental anticyclone expanded.

in a strong southwesterly current across central Canada. The sea level cyclone tracks shifted from southeast to northwest of Iowa during this period (Chart X) and the northwesterly cyclonic steering of sea level anticyclones crossing the Midwest gave way to southwesterly anticyclonic steering (Chart IX).

Thus the initiation of a continental anticyclone was favored by (1) increased west-southwesterly flow in the Pacific Northwest due essentially to momentum flux from upstream, (2) increased northerly flow along the east coast associated with the cut-off Low and westward propagation of blocking in the Atlantic, and (3) increased subtropical easterly circulation associated with easterly wave activity over the eastern Gulf of Mexico. This anticyclone, because of the seasonal heating differential between continent and ocean, then acquired dynamic characteristics (in particular, a very warm core) favoring persistence [7].

Toward the end of the third week of June, a large migratory upper-level anticyclonic cell, breaking off from the East Pacific High, entered the western United States, accompanied on its north by one of the strongest bursts of Pacific westerlies of the month. The sea level anticyclone tracks (Chart IX) show this as a retrogression of the East Pacific High from the 19th to the 21st and the formation of a new High center in Wyoming on the 21st. This shift of the jet stream to a relatively high latitude, with marked anticyclonic vorticity on its right, ended the confluence and heavy rains in the North Central States. This transition will be discussed in the next section.

THE LAST PERIOD: RIO GRANDE FLOODS, "ARIZONA RAINS," AND WIDESPREAD HEAT

In order to show more clearly the rearrangement of moist and dry air currents which took place with the ending of the confluence in the North Central United States, charts of the mixing ratio on the 700-mb. surface for 0300 GMT, June 20, June 23, and June 26 are shown in figure 7. These can be related to the 5-day mean flow patterns for June 16–20, June 19–23, and June 23–27 (fig. 6B, C, and D). Although 700-mb. charts have been used in this analysis, the resulting patterns and their associated weather conform qualitatively to those described by means of isentropic analysis in the 1930's by Namias and Wexler [7, 10].

The migratory upper-level anticyclone which entered the West on June 21 passed eastward and merged with the continental anticyclone, resulting in retrogression of the 5-day mean High (fig. 6C) from the southern Appalachians to the central Mississippi Valley. The path of this travelling anticyclone is reflected in a strong east-west mean ridge across the western United States, with

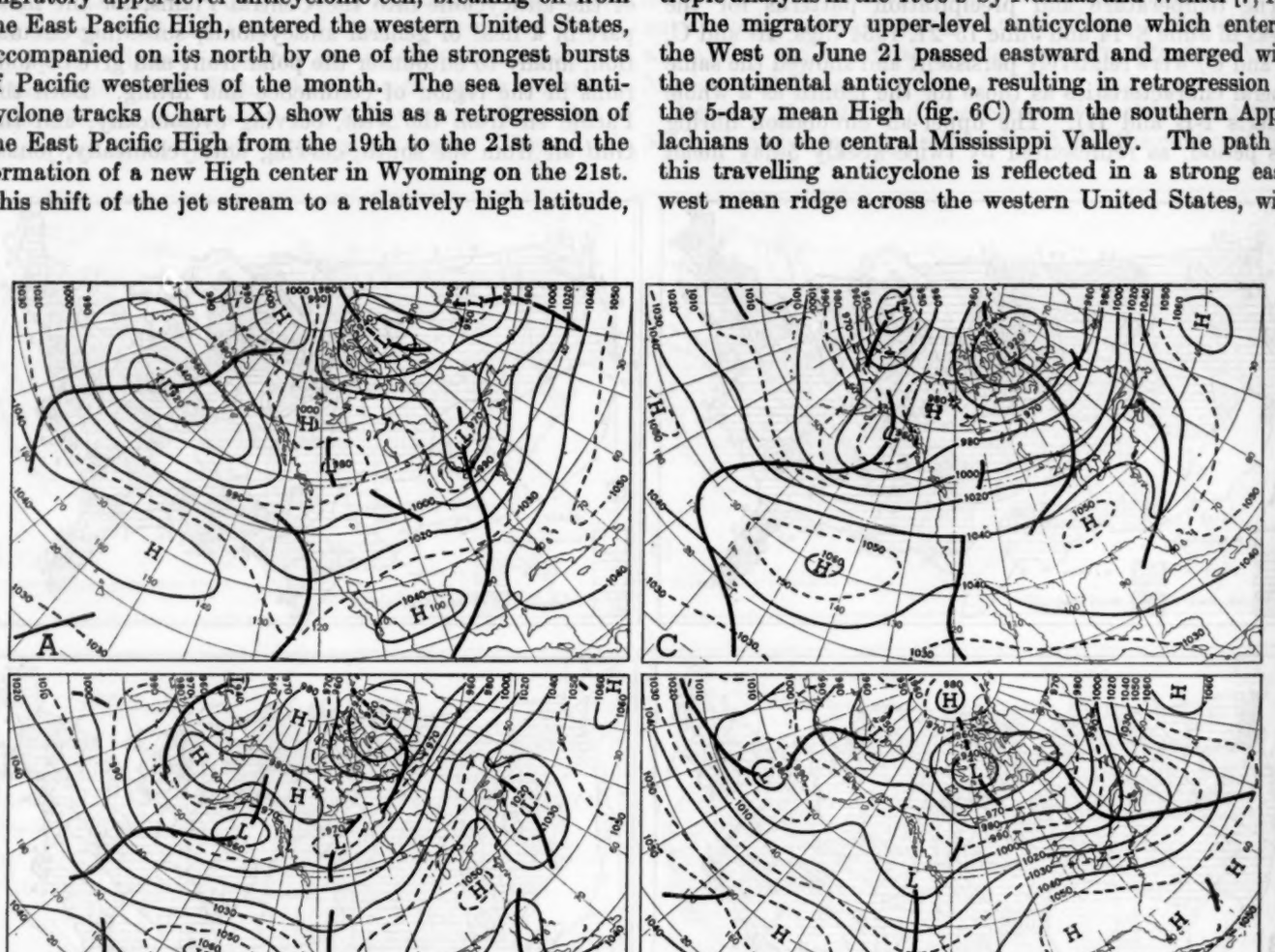

FIGURE 6.—Five-day mean 700-mb. contours (in tens of feet) for the period (A) June 2–6, (B) June 16–20, (C) June 19–23, and (D) June 23–27, 1954. A blocking surge crossed the Maritime Provinces from the east to merge with the incipient continental High in the eastern United States, while a maximum of westerly winds came across the Pacific and northern United States.

flat westerly flow on the north and flat easterly flow on the south.

Figure 7A shows the initial state of the 700-mb. moisture field preceding this intrusion of anticyclonic vorticity from the west. A moist tongue curved from eastern Mexico to the northern Great Lakes where it branched into two parts. A tongue of dry Pacific air curved cyclonically through Arizona and Colorado, while a second dry tongue covered the east coast and extended into the central Mississippi Valley. This instantaneous moisture distribution, observed at the end of the 5-day period represented by figure 6B, reflects closely the details of the mean streamlines, which had been relatively stationary during this period. To illustrate the previous histories of some of the air parcels, trajectories have been computed from the twice-daily 700-mb. charts. In constructing these trajectories it has been convenient to assume that each streamline pattern is stationary for 6 hours before and after map time and that the particles remain on the 700-mb. surface. On these assumptions, the heavy lines in figure 7 represent the travel, for the 5 days preceding the time of the mixing ratio chart, of the air parcels arriving at Sioux City, Iowa, Oklahoma City, Okla., Del Rio, Tex., and Washington, D. C. Positions of each parcel at 24-hour intervals are shown by heavy dots along the trajectories. While the particular air parcels do not, in general, remain at 700 mb., these trajectories can be presumed to represent, at least qualitatively, the antecedent travel of the main lower-tropospheric air streams arriving at these points. It can be seen that:

(1) The air at the 700-mb. level reaching northwestern Iowa, where torrential rains had fallen for the preceding three days and continued for one day following, was apparently of Pacific origin and had passed through the deep mean trough in the Southwest, curving cyclonically south of the Mexican border en route. Although not shown here, a trajectory arriving at the same time in central or southeastern Iowa was found to have come anticyclonically from the Gulf of Mexico. In other words, the histories of the two air streams arriving in Iowa on June 20 are fairly well represented by the 10,300-ft. and 10,400-ft. contours on the 5-day mean 700-mb. chart for June 16–20. The 6 gm/kg mixing ratio line roughly separates these two streams. Of course the actual values cannot be explained entirely by horizontal advection since Gulf moisture was undoubtedly being carried upward from lower levels.

(2) Both the dry air arriving at Oklahoma City and the moist air arriving at Del Rio at the 700-mb. level had come from over warm tropical waters. While the trajectories show no pronounced curvature, both were under the influence of a broad anticyclonic flow field during their 5-day travel. No significant rain was reported from either station on this date or for several days before or after.

(3) The dry cool air arriving at Washington, D. C., from

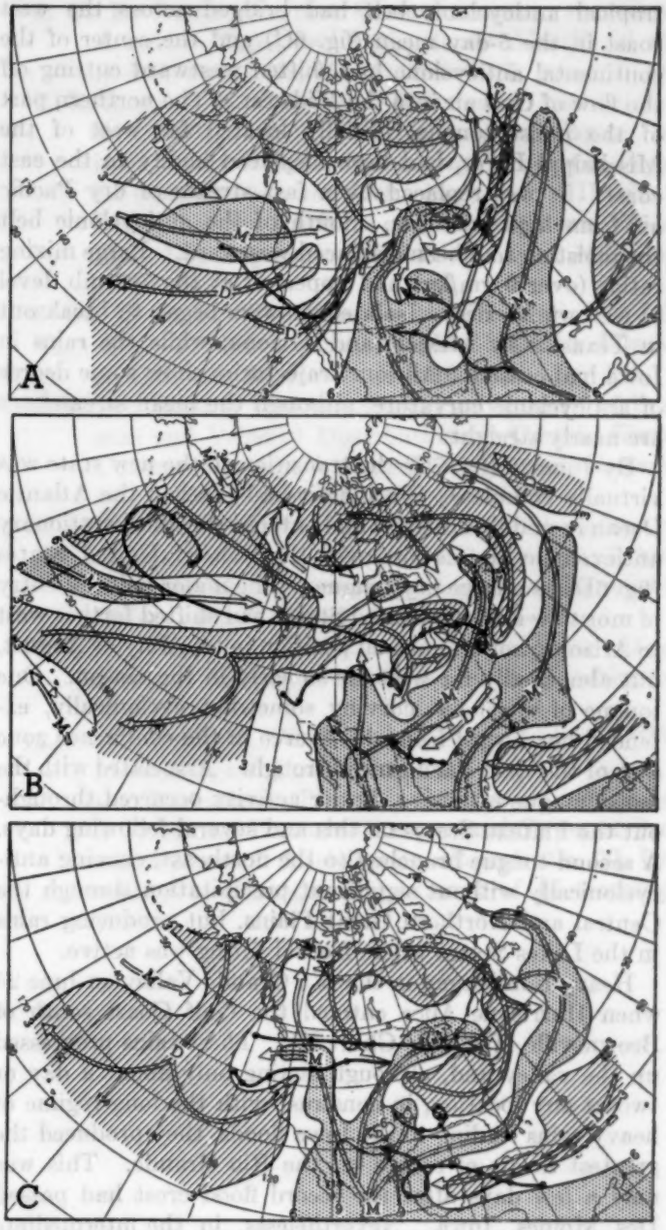

FIGURE 7.—700-mb. mixing ratio (gm/kg) at 0300 GMT for (A) June 20, (B) June 23, and (C) June 26, 1954. Areas of less than 3 gm/kg are hatched and areas of more than 6 gm/kg are stippled. Axes of relatively moist (M) and dry (D) are shown by double arrows. Trajectories for the five days preceding map time are shown as heavy arrows for air parcels arriving at Del Rio, Tex., Oklahoma City, Okla., Sioux City, Iowa, and Washington, D. C.; heavy dots mark 24-hour intervals along the trajectories. Moisture initially over the Great Plains was displaced by fast westerlies but a new supply appeared farther west when the anticyclone stagnated over the central Mississippi Valley.

the north on June 20 could apparently be traced back directly to the southern tip of Greenland 5 days earlier.

Thus qualitatively, toward the end of this period of relatively stable flow, the moisture axes (lines of relative maximum and minimum mixing ratio at map time), streamlines, and trajectories all told essentially the same story. Three days later (fig. 7B), however, some very striking changes had occurred, and the three sets of lines were no longer in accord. As we have seen, the sub-

tropical anticyclonic belt had bridged across the west coast in the 5-day mean (fig. 6C) and the center of the continental anticyclone had shifted westward cutting off the flow of Gulf air into the Midwest. The northern part of the moist tongue, initially located far west of the Mississippi River, had been displaced bodily to the east coast. It was replaced by a fast stream of dry Pacific air from high latitudes. South of the anticyclonic belt the moist air had been displaced westward. Large mixing ratios (over 9 gm/kg) had appeared at the 700-mb. level in western Mexico. Isolated showers began to break out in Texas, New Mexico, and Arizona, while the rains in Iowa had ceased. All four trajectories show some degree of anticyclonic curvature, although the mean streamlines are nearly straight.

By June 26 (fig. 7C) the transition to the new state was virtually complete. Pacific air had reached the Atlantic Ocean in middle latitudes and entered the quasi-stationary anticyclonic circulation over the eastern United States (fig. 6D) as a large dry tongue. The major point of entry of moist air into the United States had shifted farther west to Arizona, not an unusual position in late July or August, but abnormally far west for so early in the season. One tongue of moist air, curving somewhat cyclonically, extended northward from this source in the confluence zone east of the East Pacific mean trough. Associated with the moist tongue extensive shower activity occurred throughout the Plateau States on this and several following days. A second tongue branched to the northeast, curving anticyclonically without significant precipitation through the Central and Northern Great Plains, but producing rains in the Lakes States where frontal lifting was active.

Heavy rains began in the Rio Grande Valley on June 25 when Hurricane Alice entered the Gulf Coast south of Brownsville, Tex. (see Chart X). These rains progressed up the valley and, although lasting only about a day or two at any location, in constrast with the long regime of heavy rains leading to the Iowa floods, they produced the greatest floods of record on the Rio Grande. This was only a few days after the record flood crest had passed Des Moines, Iowa. Nevertheless, in the intermediate area of Kansas, Oklahoma, Missouri, and Arkansas no significant precipitation occurred during the entire last half of June. It is of interest that the mixing ratio at 700 mb. and lower levels was nearly the same at Del Rio and Oklahoma City on June 26. However, it is clear from the trajectories in figure 7C that the air arriving at Oklahoma City had travelled anticyclonically from near the center of the continental High, while that arriving at Del Rio had been caught in the cyclonic circulation around the tropical storm. This difference was, however, reflected in the moisture content at 500 mb., where the mixing ratio was 3.7 gm/kg at Del Rio and only about 1 gm/kg at Oklahoma City. This suggests that the vertical motion and moisture supply through a considerable layer above the 700-mb. surface, as affected by the field of motion along the path, played a part in producing drought at one place and floods at the other. On June 28 the Texas rains were largely ended as the deep moist air over southern Texas was replaced by a tongue of dry air which can be seen progressing steadily westward in the three charts of figure 7.

During the fourth week of June daily maximum temperatures rose to over 100° F., first in Kansas and then in nearly all of the United States south of 40° N. New all-time high temperatures were recorded at Casper, Wyo., and Macon, Ga. Extreme heat is one of the essential characteristics of the quasi-stationary high-level continental anticyclone when well developed. As a result, the sea level Highs were weak in this area; the 5-day mean sea level charts showed high centers of about 1,017 mb. in the Southeast as contrasted with over 1,025 mb. in both of the oceanic subtropical Highs. Yet at 700 mb. the continental High was just about as strong as the oceanic subtropical Highs, while at 200 mb. the 5-day mean High of over 40,800 ft. over Kansas and Oklahoma for June 23–27 was the highest center in the Western Hemisphere. The large continental High with troughs over the oceans on both sides is a common summertime feature at 10–19 km. This was shown both by the high-level normal charts constructed by the Weather Bureau in 1944 [11] and through the study and typing of daily high-level charts by Wulf, Hodges, and Obloy [12].

SUMMARY

During the month the birth, growth, and retrogression of the continental anticyclone in a setting of high index over the Pacific and low index over the Atlantic was reflected in the temperature, moisture, and precipitation patterns as follows:

(1) *Temperature.*—Unseasonably cold weather, persisting from May and covering almost the entire country in the first week but with particularly severe frosts in the Great Basin, was replaced by unseasonable warmth, appearing first in the Mississippi-Ohio Valley and later spreading over most of the rest of the country to reach heat-wave proportions over large areas. Exceptions were the extreme Pacific Northwest which was dominated by cold Pacific air throughout the month, and the East which experienced repeated prolonged invasions of cool Atlantic air.

(2) *Moisture.*—The major tropical moist tongue entering the middle latitudes, at first virtually confined to the extreme southeastern States, came northward over the central part of the country during the middle of the month and shifted to the western Plateau States toward the end. Dry tongues from the Pacific, at first covering much of the country, retreated until they merely skirted the western and northern borders. Dry tongues entered the East at first directly from the North Atlantic and later from the Pacific, returning anticyclonically.

(3) *Precipitation.*—Frontal lifting was the predominant cause of precipitation in the first week with significant amounts occurring throughout most of the northeastern half of the country. Heavy rains also occurred in tropical air over Florida. During the middle of the month the most spectacular rains occurred in the North but drought began to develop in the central Mississippi Valley and parts of the Central Plains. The final third of the month brought showers to the Southwest and Western Plateau, with drought spreading over the Midwest.

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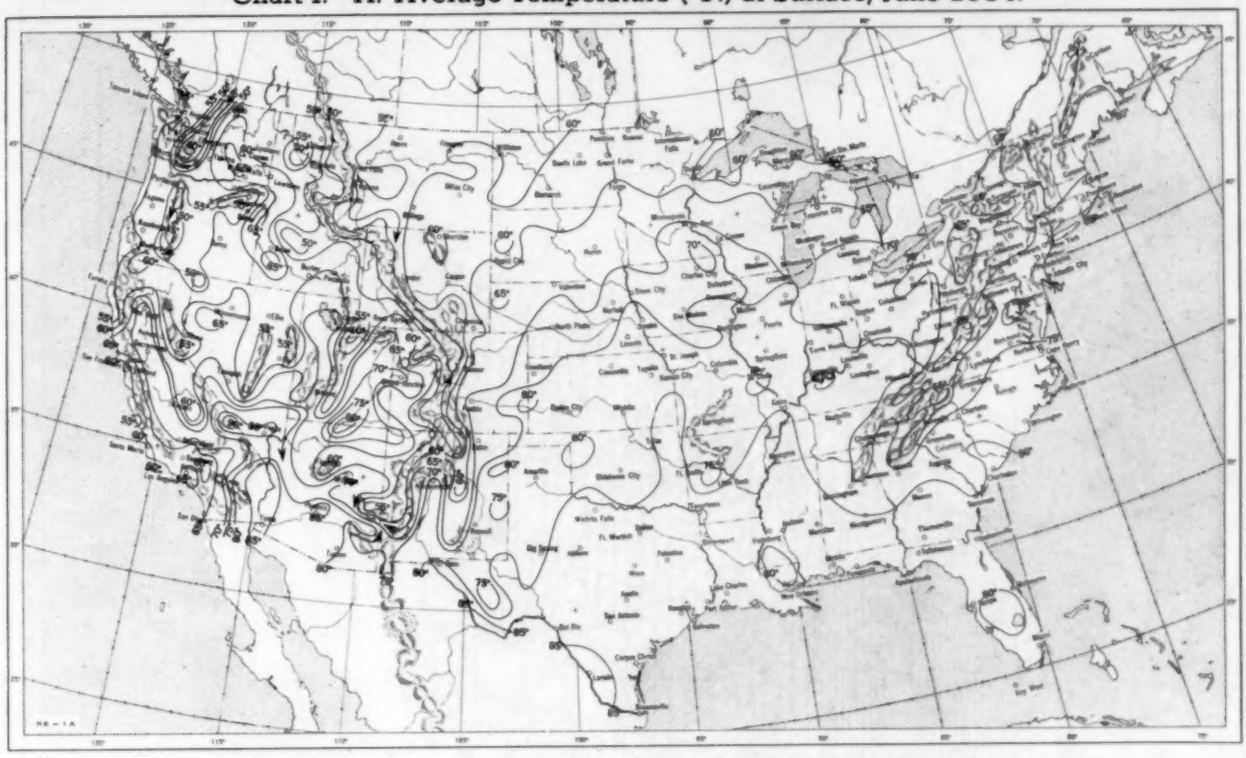
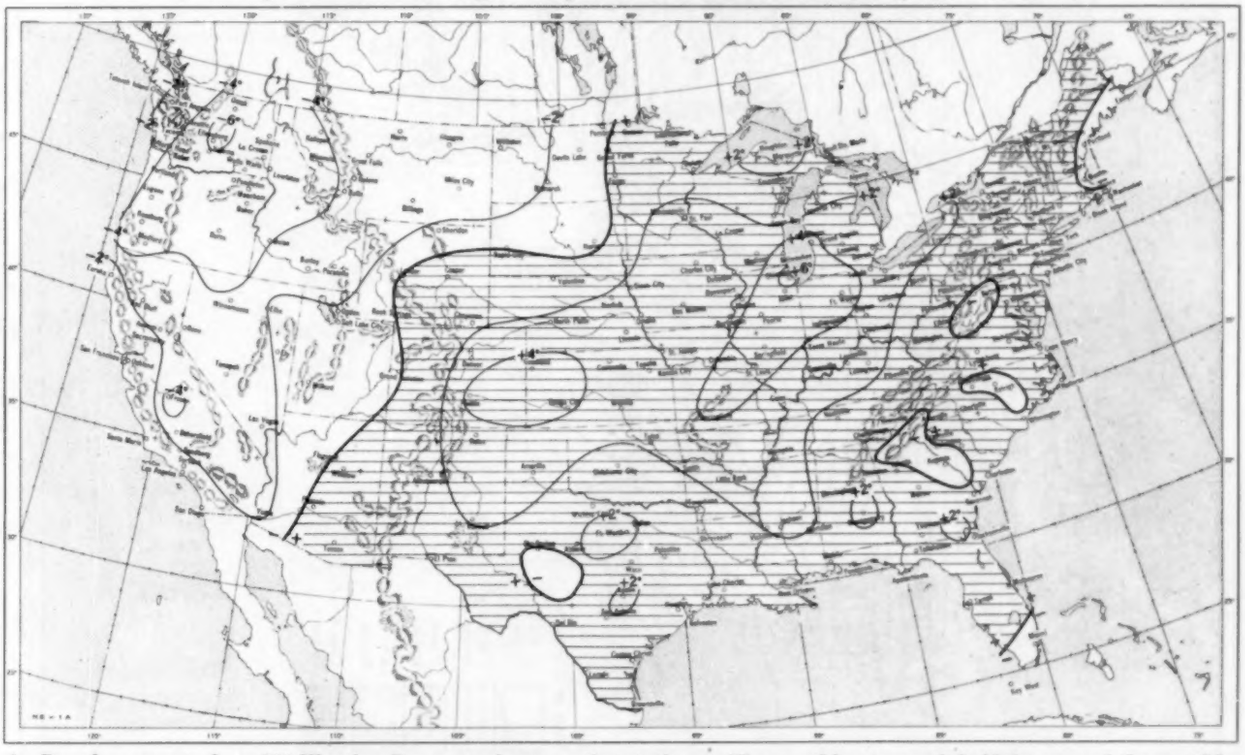
CORRECTION

MONTHLY WEATHER REVIEW, vol. 82, Nos. 3 and 4: Charts IX and X, Tracks of Anticyclones and Cyclones for March 1954 were interchanged with Charts IX and X for April 1954. The tracks printed in the March issue are those for April; the March tracks appear in the April issue.

Chart I. A. Average Temperature (°F.) at Surface, June 1954.

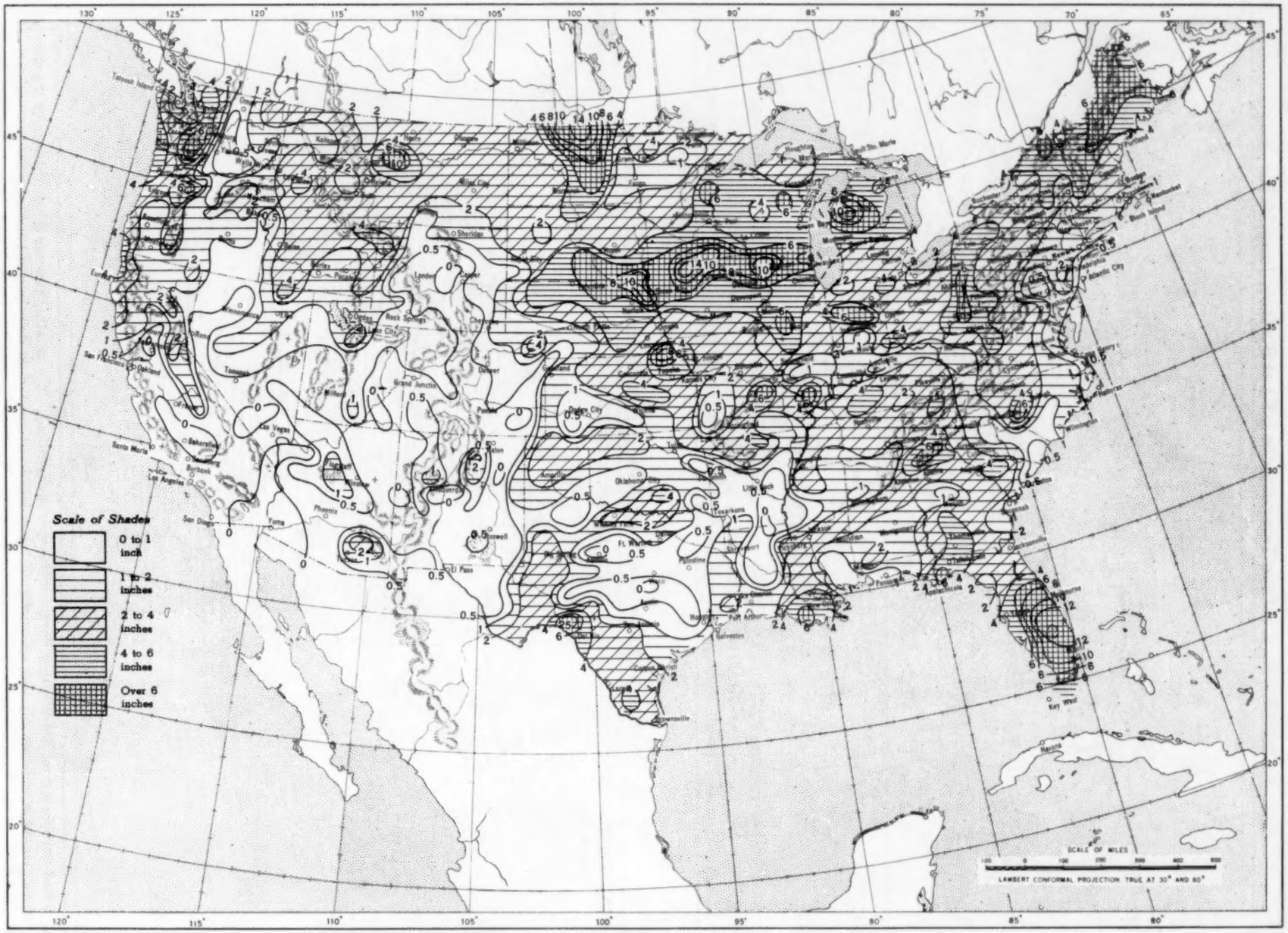


B. Departure of Average Temperature from Normal (°F.), June 1954.



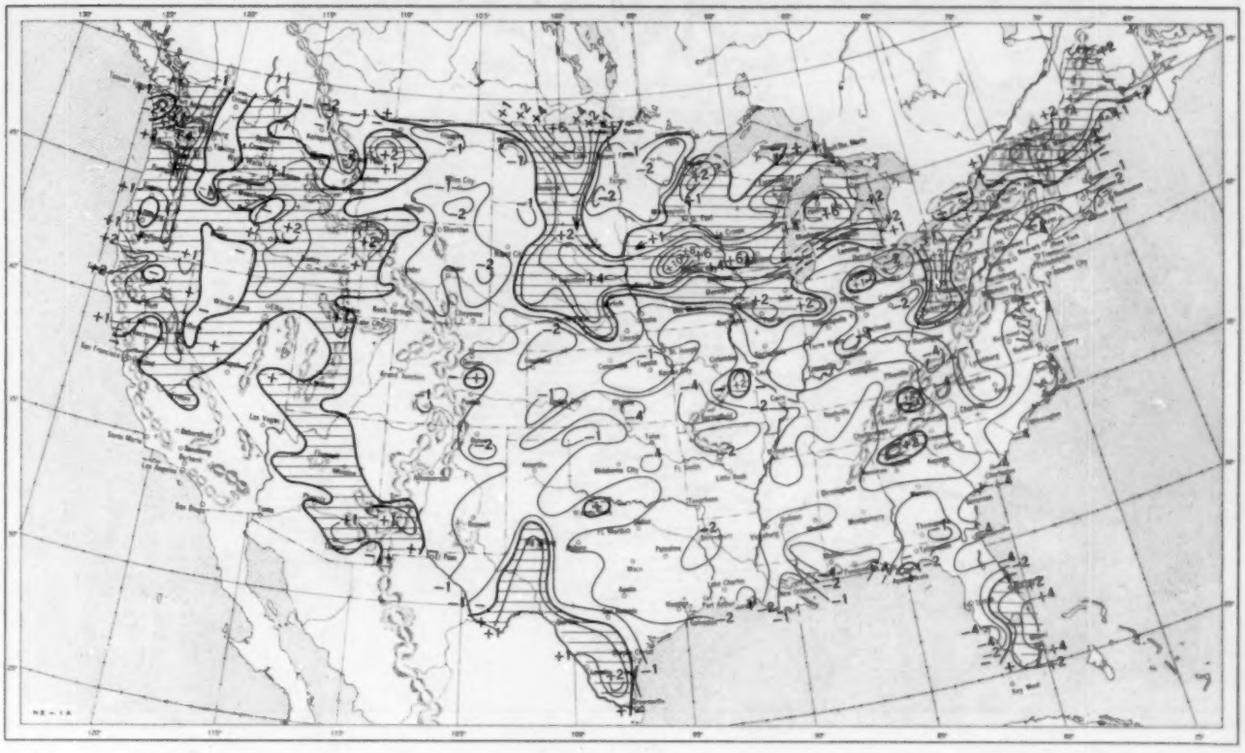
A. Based on reports from 800 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.
B. Normal average monthly temperatures are computed for Weather Bureau stations having at least 10 years of record.

Chart II. Total Precipitation (Inches), June 1954.

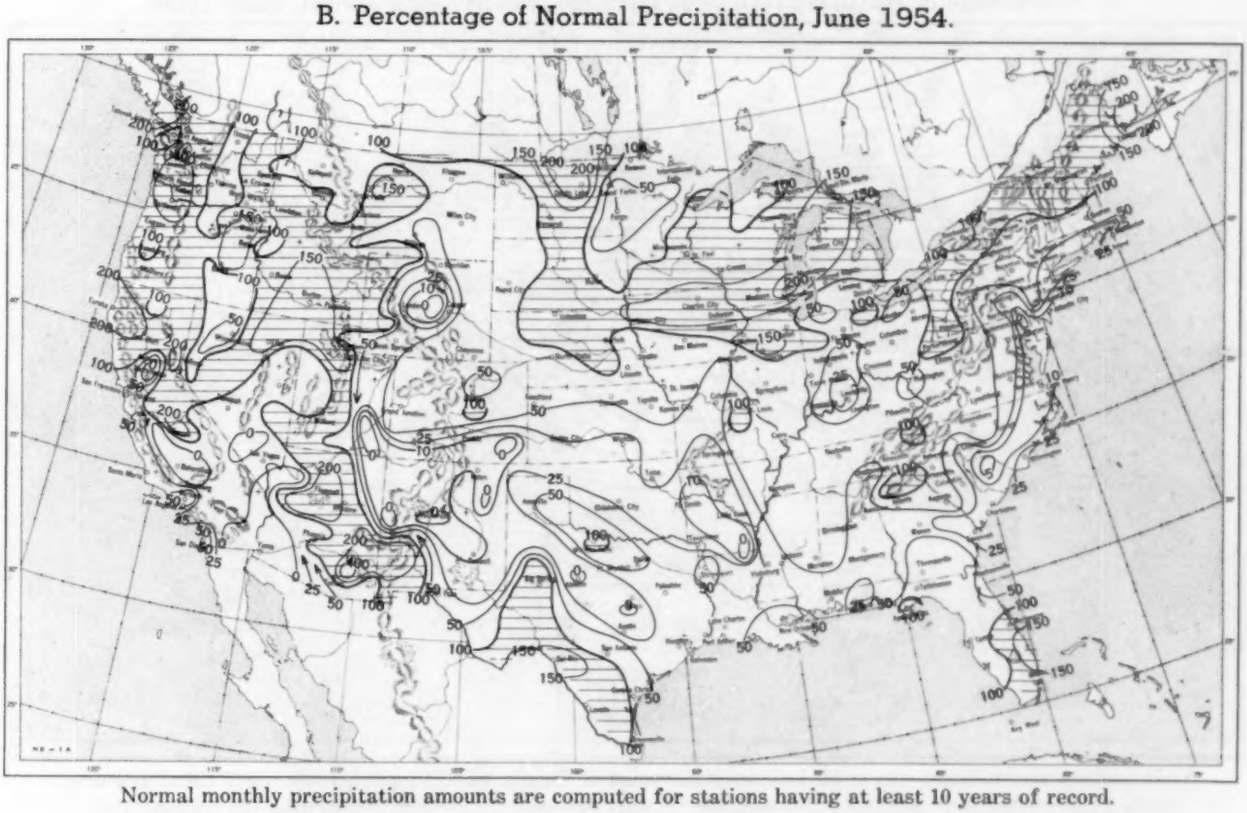

Scale of Shades
0 to 1 inch
1 to 2 inches
2 to 4 inches
4 to 6 inches
Over 6 inches
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°

Based on daily precipitation records at 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), June 1954.

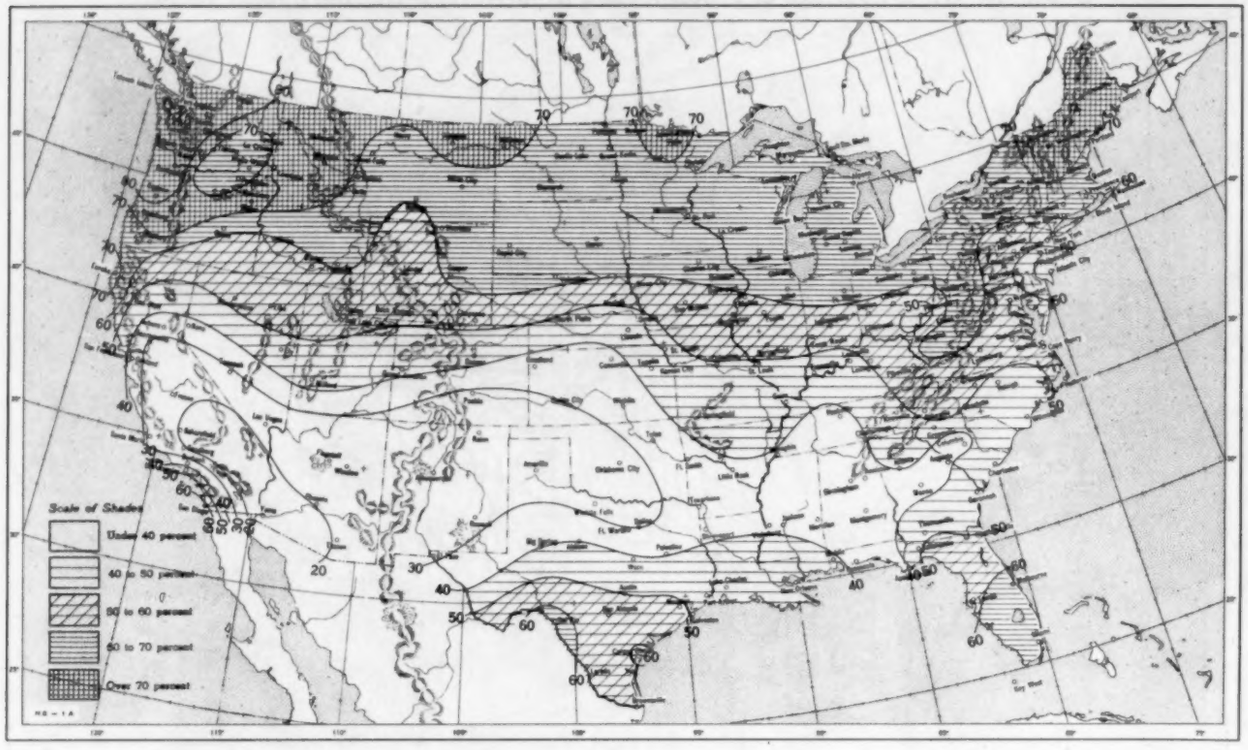


B. Percentage of Normal Precipitation, June 1954.

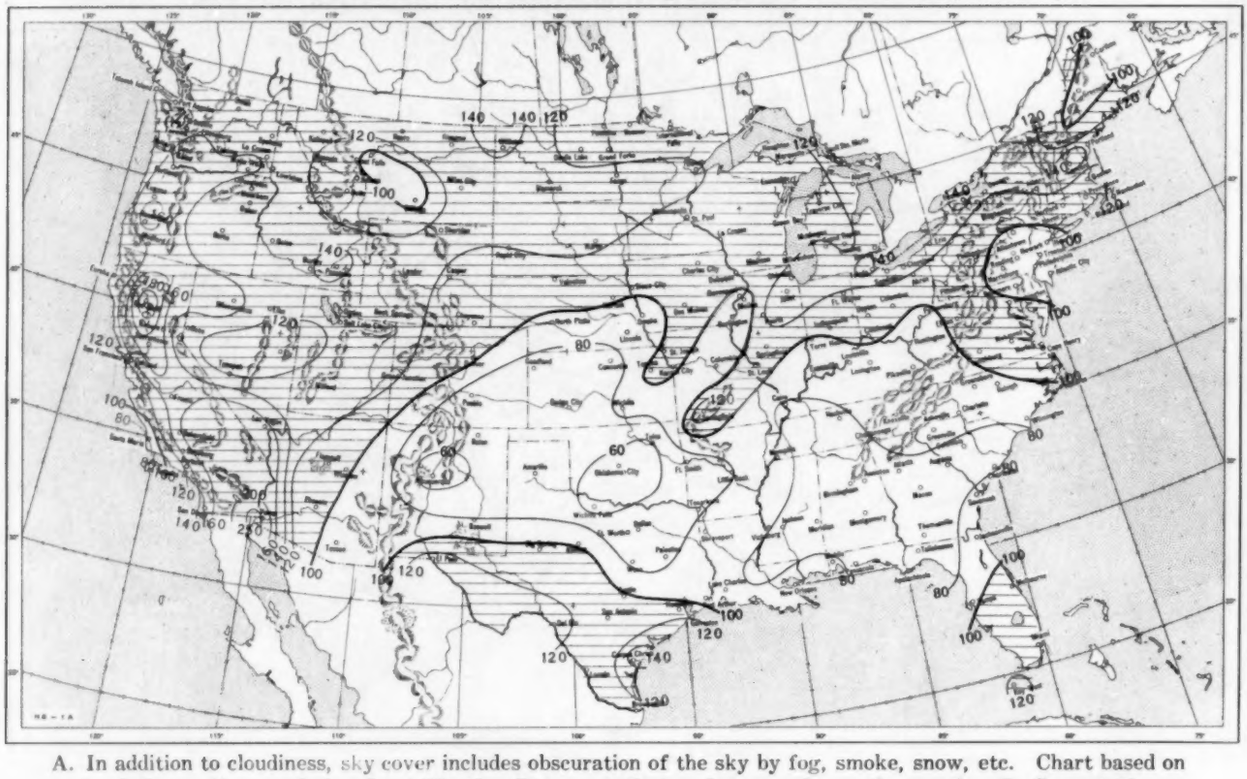


Normal monthly precipitation amounts are computed for stations having at least 10 years of record.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, June 1954.

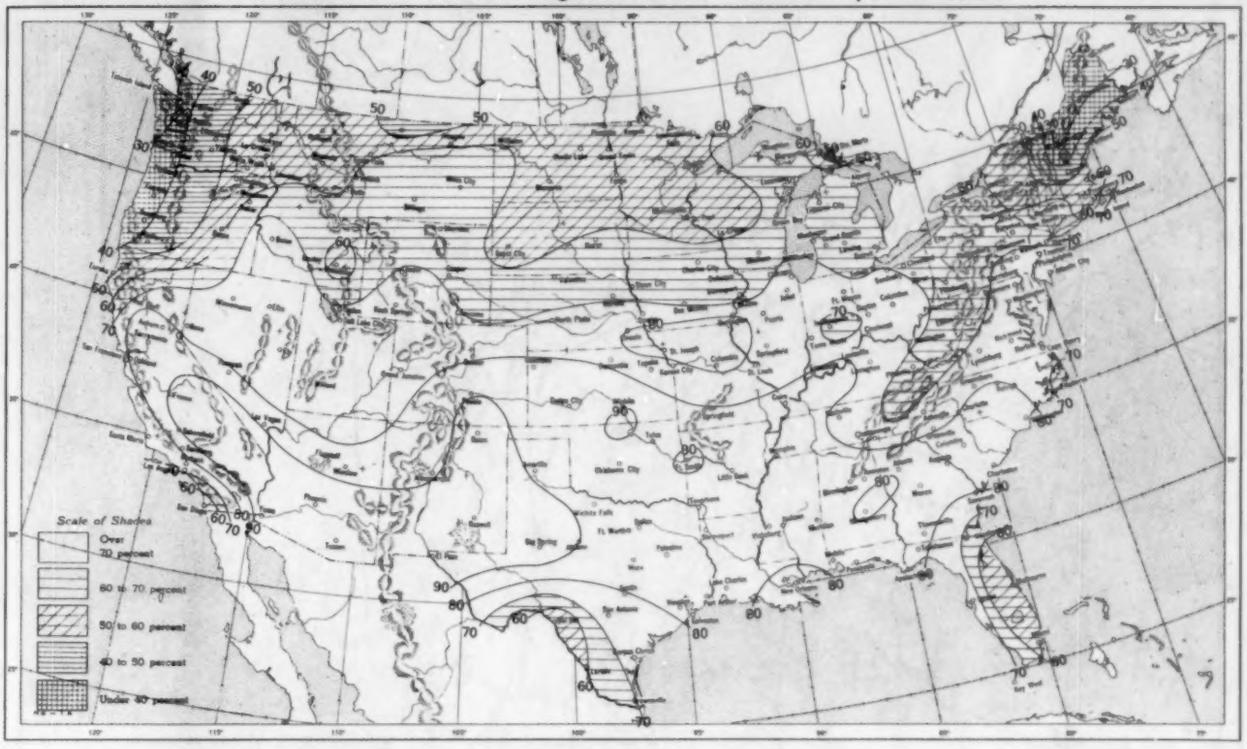


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, June 1954.

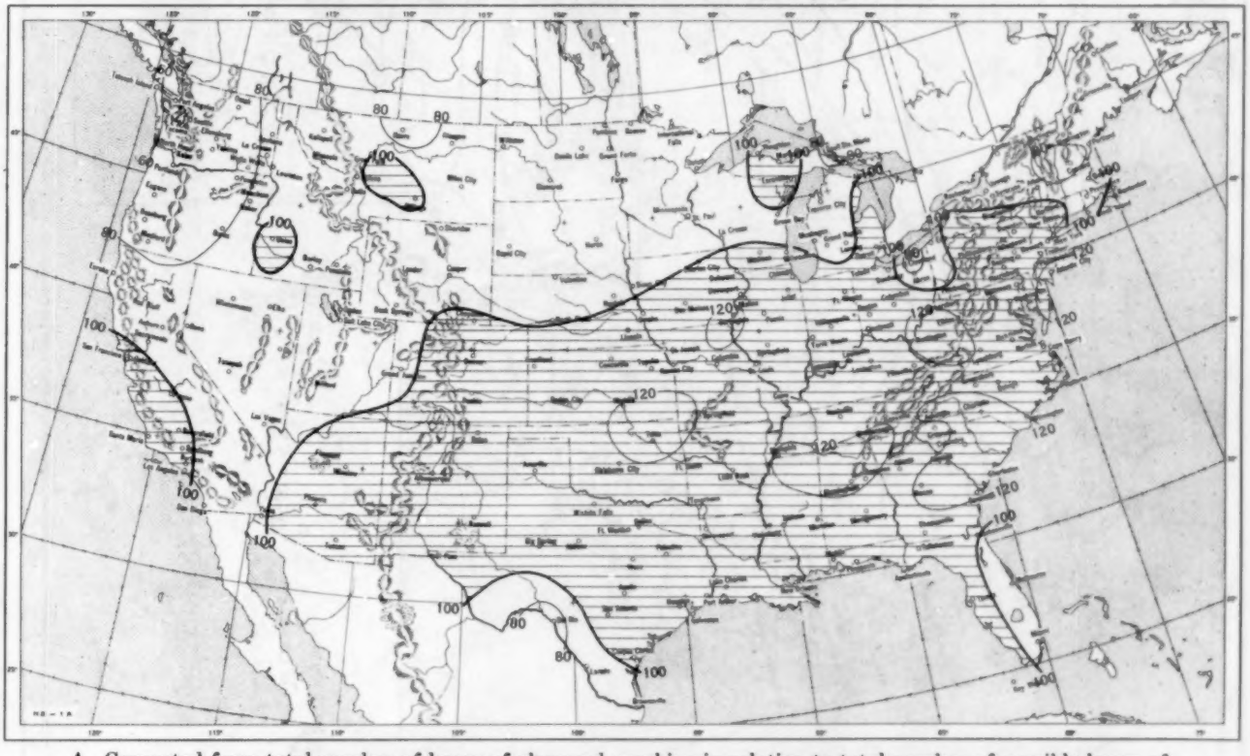


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, June 1954.



B. Percentage of Normal Sunshine, June 1954.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, June 1954. Inset: Percentage of Normal Average Daily Solar Radiation, June 1954.

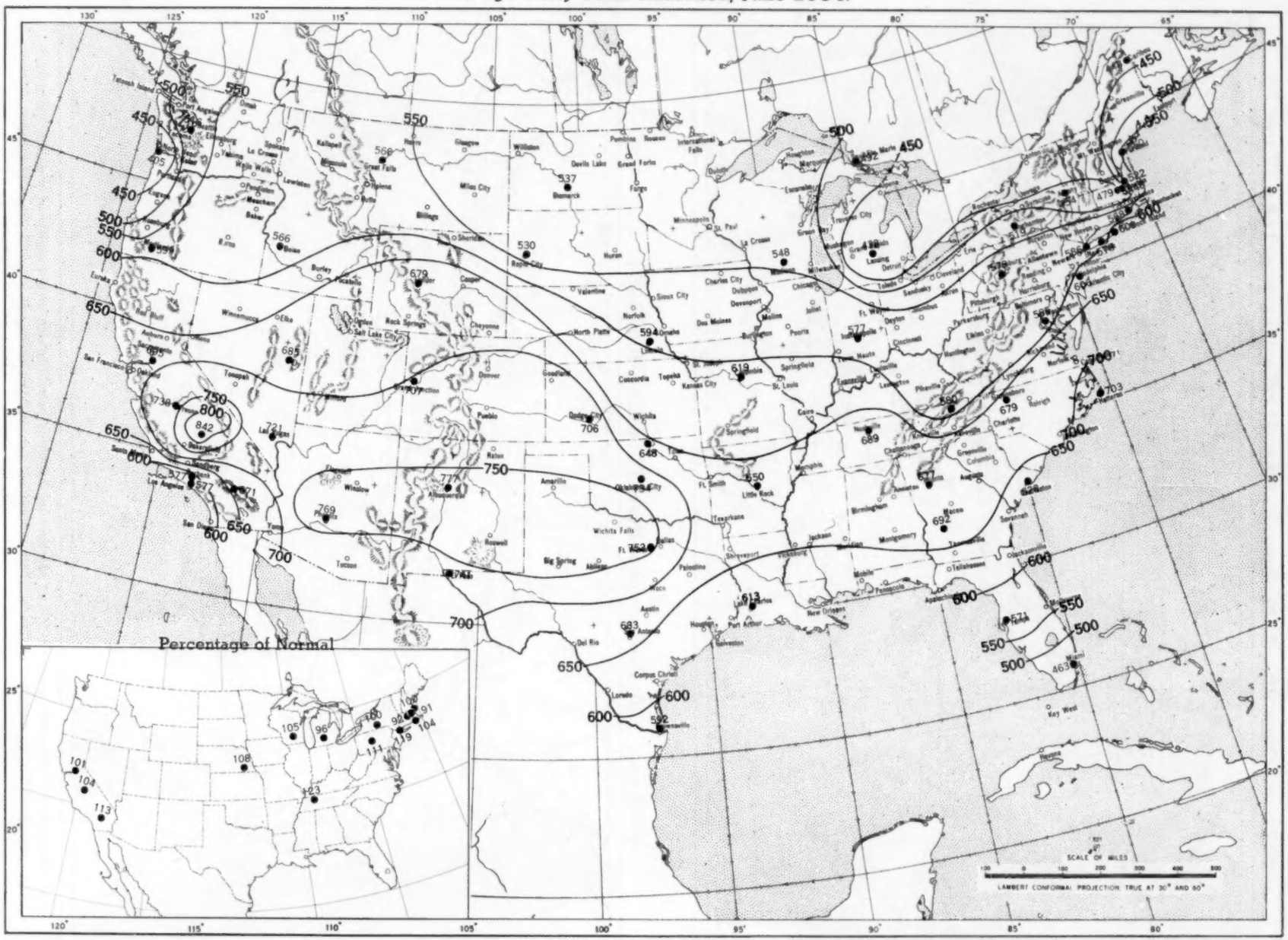
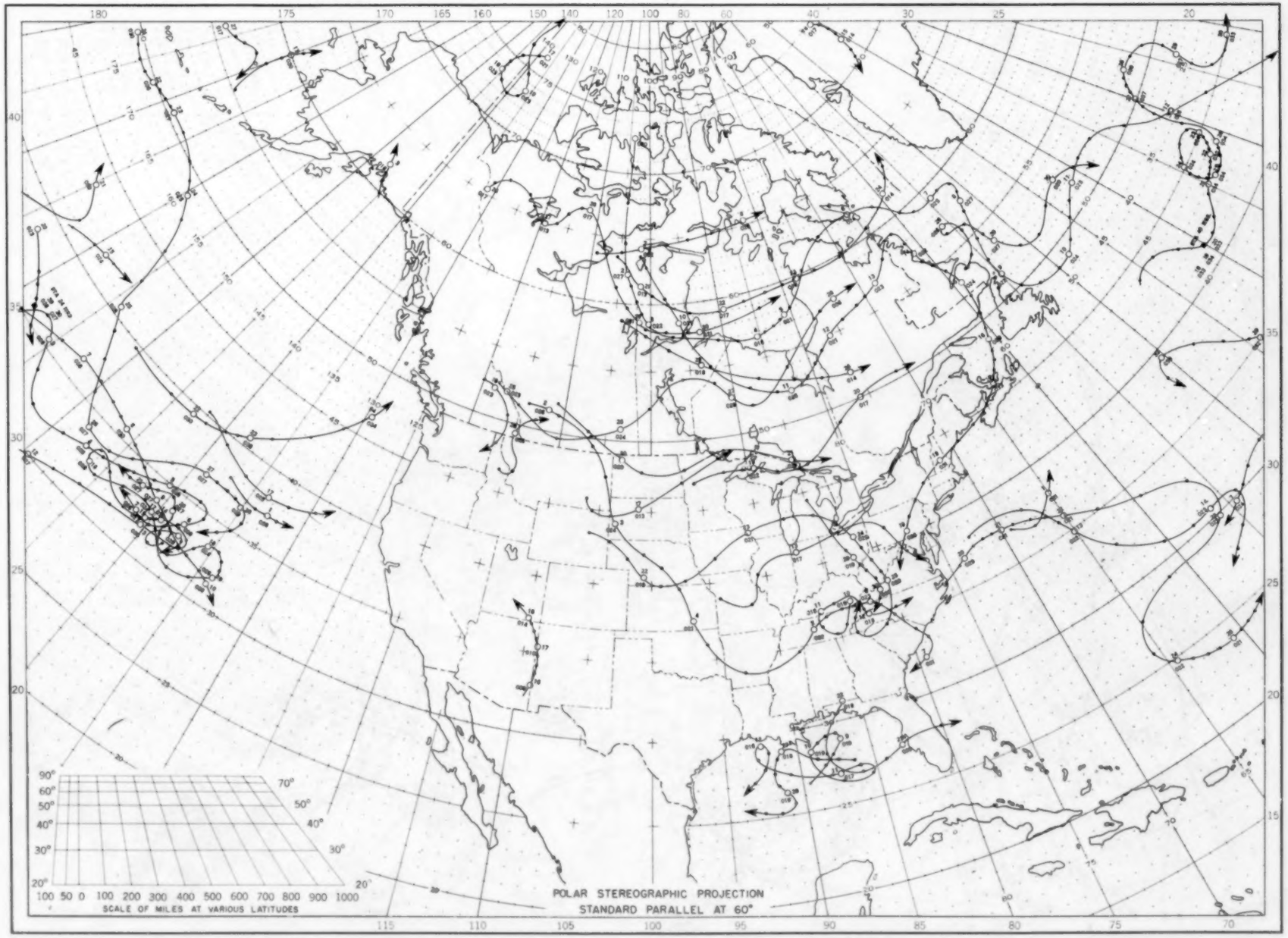


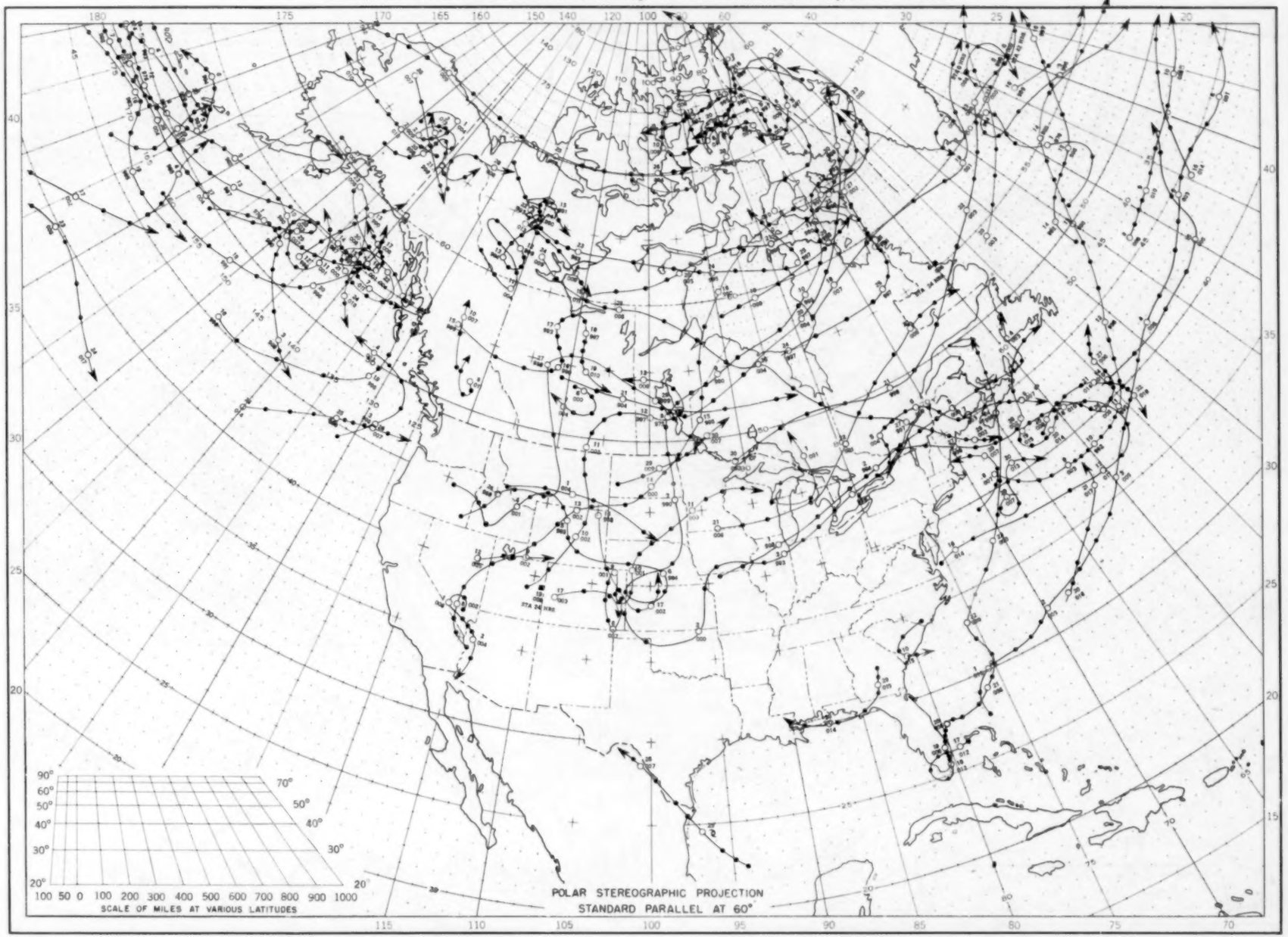
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown. Normals are computed for stations having at least 9 years of record.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, June 1954.


SCALE OF MILES AT VARIOUS LATITUDES
100 50 0 100 200 300 400 500 600 700 800 900 1000
POLAR STEREOGRAPHIC PROJECTION
STANDARD PARALLEL AT 60°

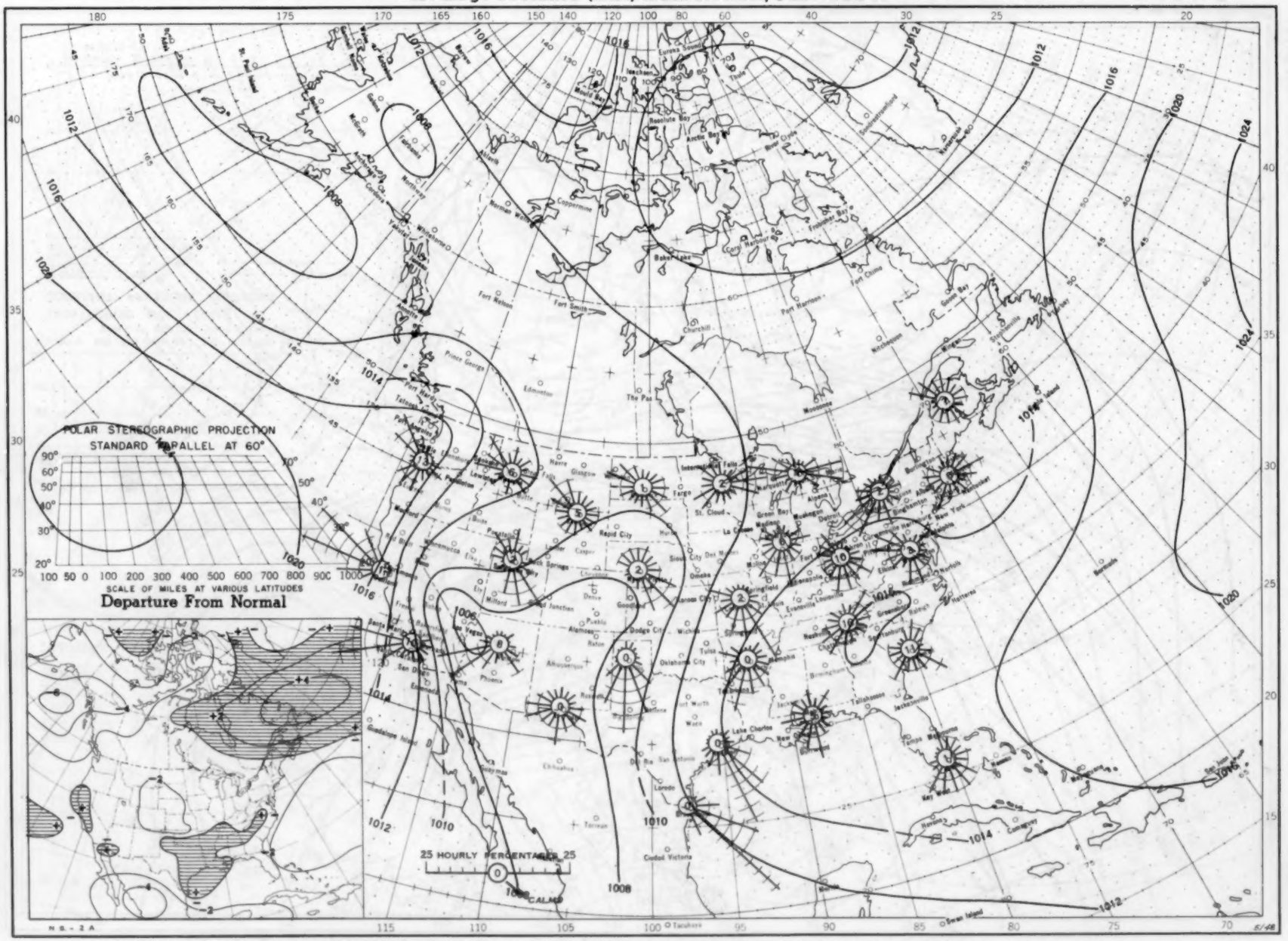
Circle indicates position of center at 7:30 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, June 1954.


90° 60° 50° 40° 30° 20°
70° 50° 40° 30° 20°
100 50 0 100 200 300 400 500 600 700 800 900 1000
SCALE OF MILES AT VARIOUS LATITUDES
POLAR STEREOGRAPHIC PROJECTION
STANDARD PARALLEL AT 60°

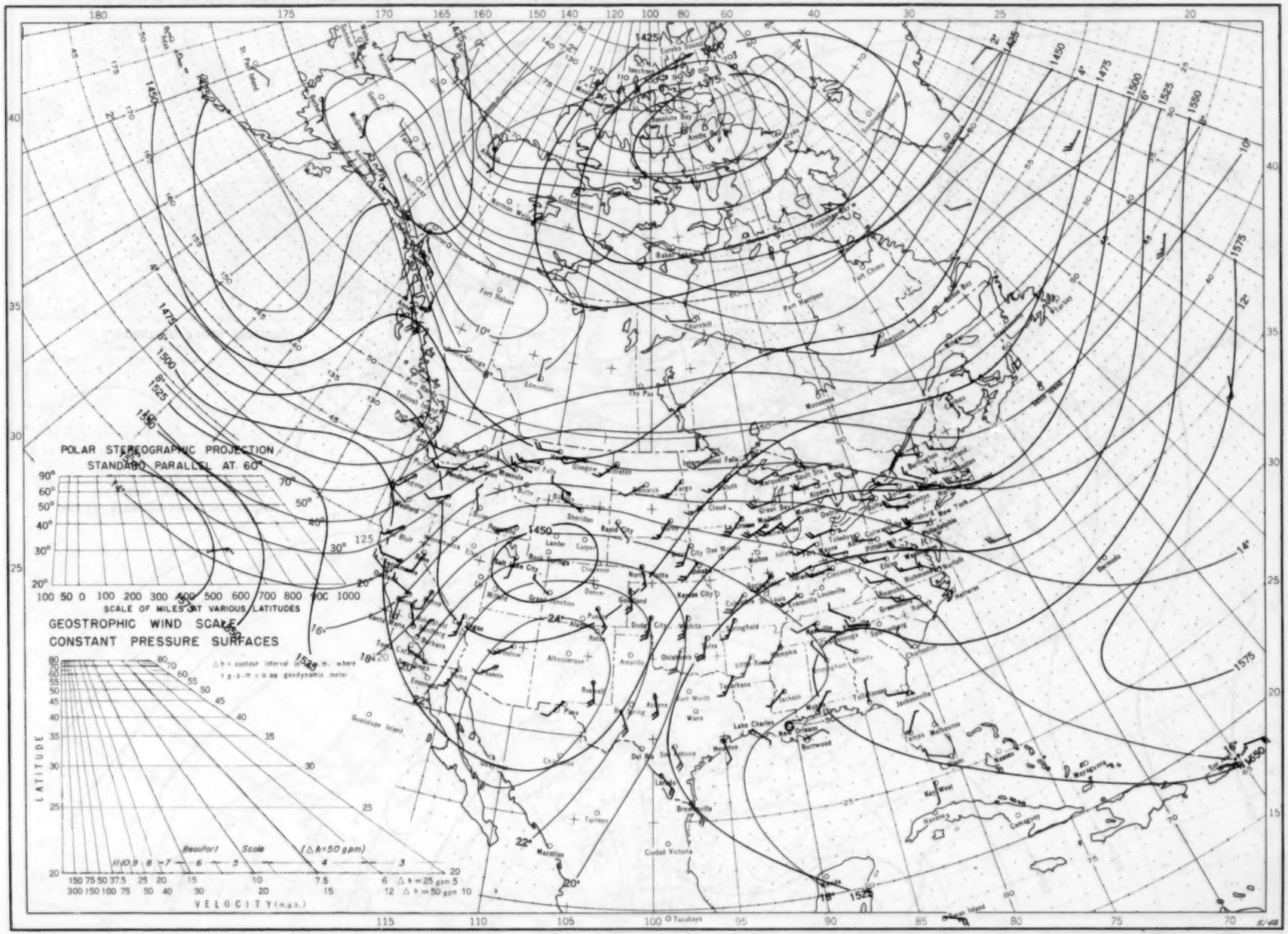
Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, June 1954. Inset: Departure of Average Pressure (mb.) from Normal, June 1954.



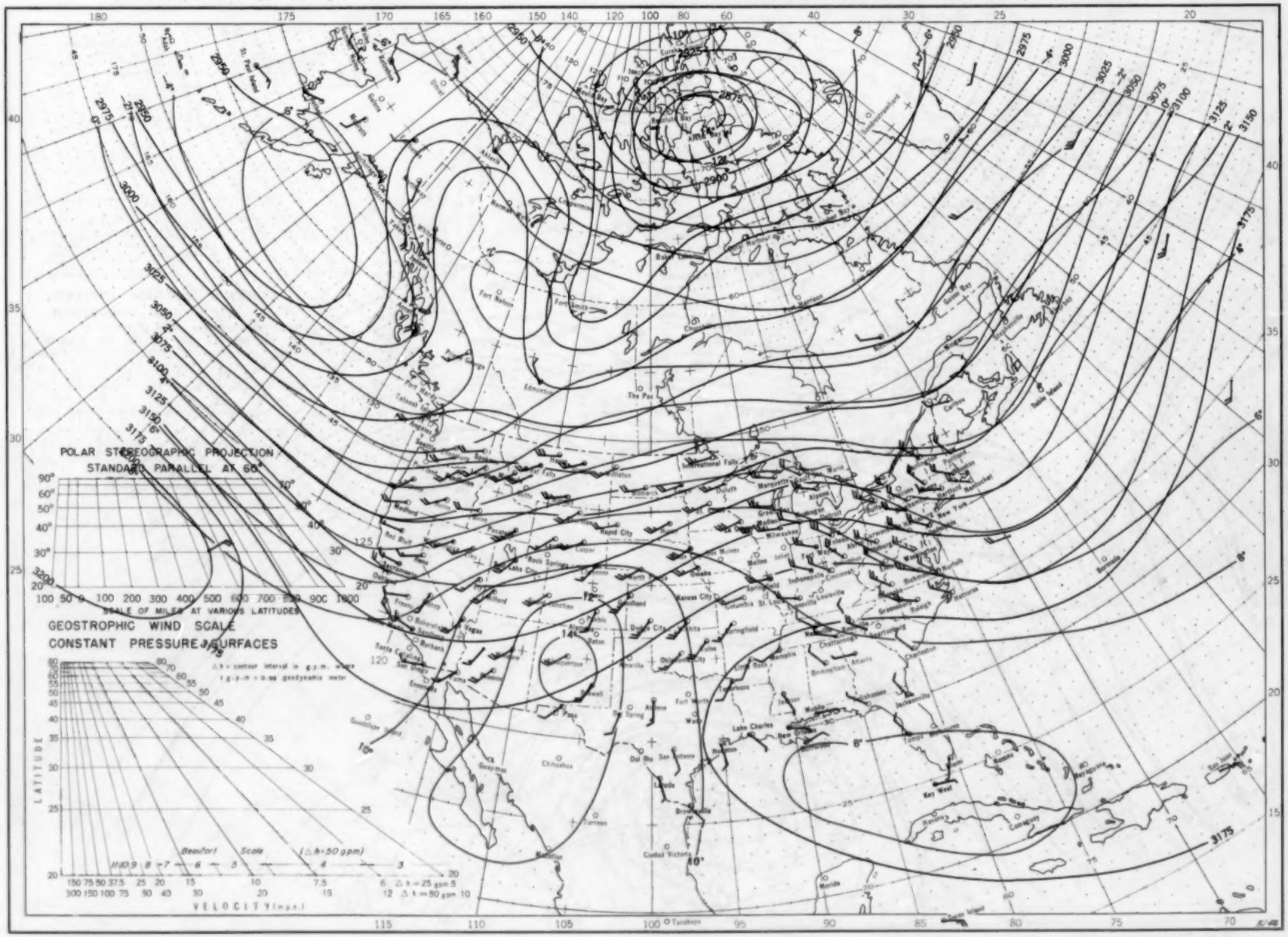
Average sea level pressures are obtained from the averages of the 7:30 a.m. and 7:30 p.m. E.S.T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 850-mb. Pressure Surface, Average Temperature in °C. at 850 mb., and Resultant Winds at 1500 Meters (m.s.l.), June 1954.



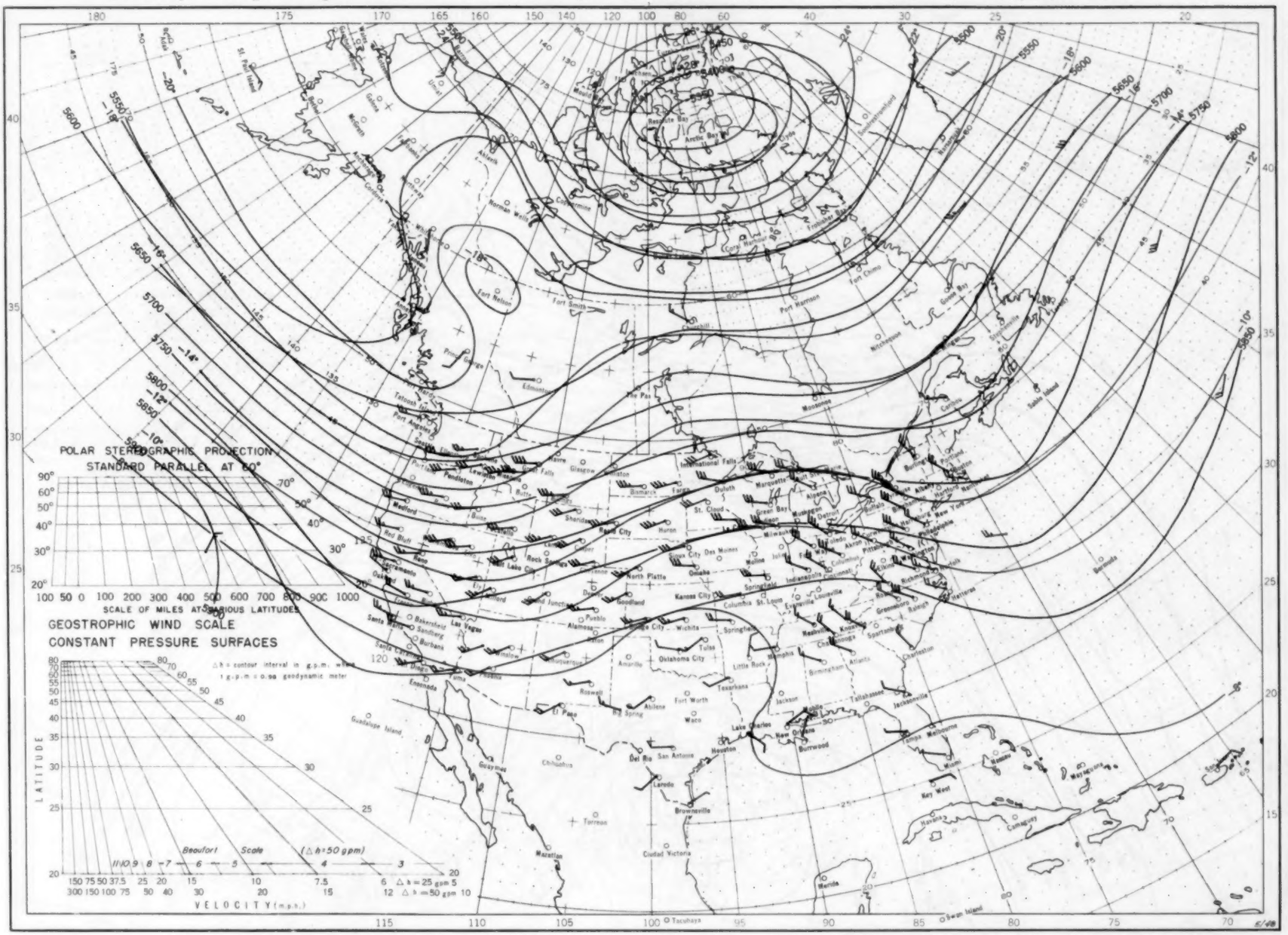
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XIII. Average Dynamic Height in Geopotential Meters (1 g.p.m.. = 0.98 dynamic meters) of the 700-mb. Pressure Surface, Average Temperature in °C. at 700 mb., and Resultant Winds at 3000 Meters (m.s.l.), June 1954.



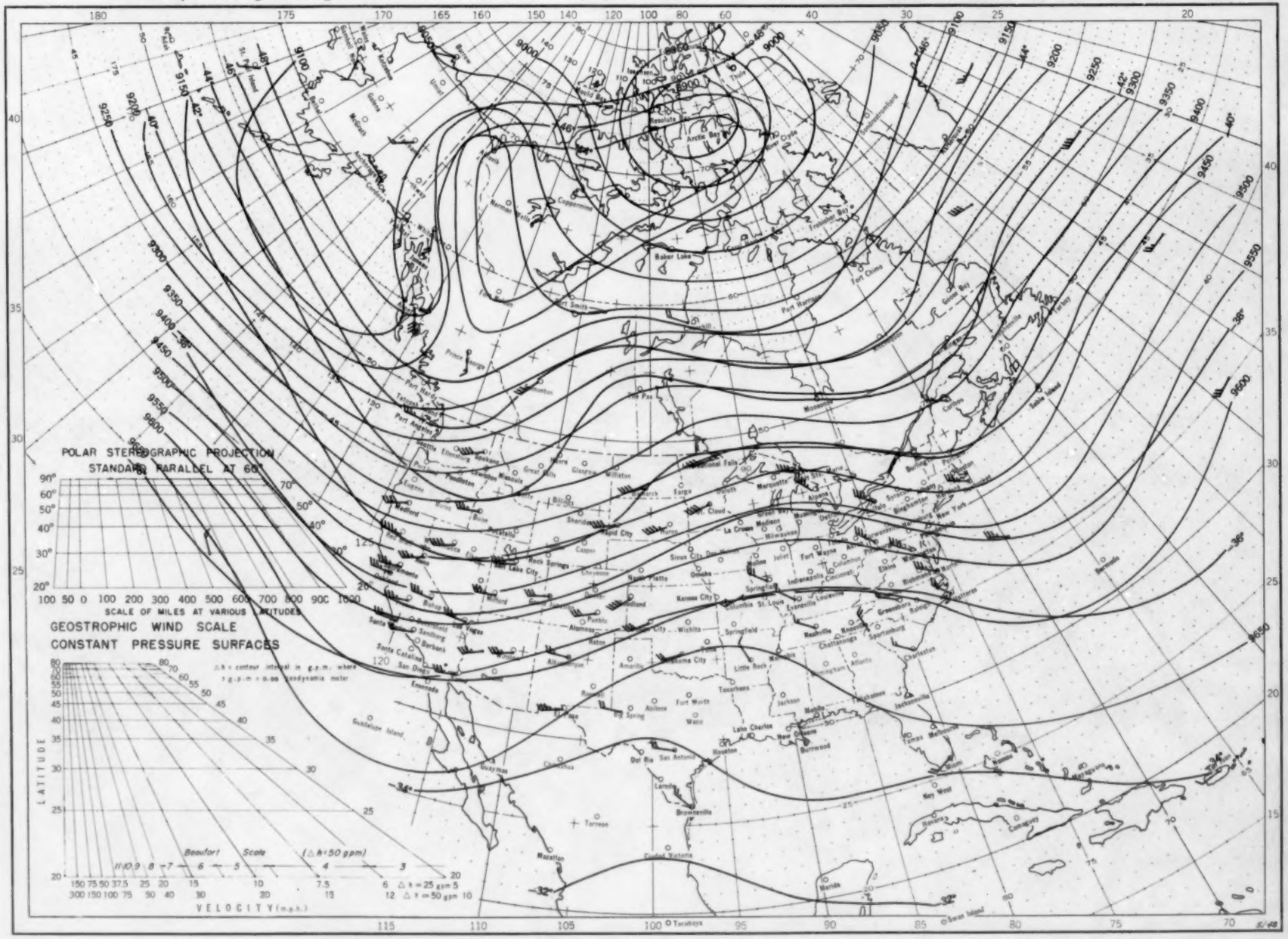
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XIV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 500-mb. Pressure Surface, Average Temperature in °C. at 500 mb., and Resultant Winds at 5000 Meters (m.s.l.), June 1954.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.

Chart XV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 300-mb. Pressure Surface, Average Temperature in °C. at 300 mb., and Resultant Winds at 10,000 Meters (m.s.l.), June 1954.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T. Wind barbs indicate wind speed on the Beaufort scale.